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Boris S. Sokolov

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Sr Isotopic Composition in Riphean, Vendian, and Lower Cambrian Carbonates from Siberia

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Abstract – Initial $^{87}Sr/^{86}Sr$ ratios were measured in 51 carbonate rock samples of Riphean, Vendian, and Early Cambrian ages from Siberian sections (Olenek, Turukhansk, and Ura uplifts). According to normal geochemical criteria, the studied rocks seem unaffected by any significant postdepositional alterations. However, all the samples contain a recrystallized carbonate phase that formed by catagenesis and differs from the early diagenetic phase in its Sr isotopic composition. Therefore, published $^{87}Sr/^{86}Sr$ values of Precambrian carbonates contain, at least partially, a systematic error that is dependent on the differences in the Sr concentrations and $^{87}Sr/^{86}Sr$ ratios of both phases. Using a new approach to separate these phases, we measured initial $^{87}Sr/^{86}Sr$ ratios in carbonate successions of the Olenek (Lower and lower Middle Riphean), Turukhansk (upper Middle and lower Upper Riphean), and Ura (uppermost Riphean, Vendian, and Atdabanian Stage of the Lower Cambrian) uplifts. Initial ratios in the Middle and lower Upper Riphean rocks range from 0.7052 to 0.7062 and are well below the values accepted in modern publications. Samples from the uppermost Riphean, Vendian, and Atdabanian successions of Siberia revealed remarkably higher values corresponding to 0.7081 - 0.7110, 0.7084 - 0.7085, and 0.7081 - 0.7085, respectively. New data adds significant detail to the Riphean–Vendian curve of seawater Sr isotopic variations. The inversion points on the curve are correlative with certain Late Proterozoic events.

INTRODUCTION

The Sr isotopic composition of seawater is dependent on the global balance between the following three material inputs into oceans: (1) continental runoff (Fc), comprising river discharge and groundwater runout, which are both averaging characteristics of the continental crust; (2) mantle flux (Fm), originating from the oceanic basalts–seawater interaction in hydrothermal convection cells of mid-ocean ridges; and (3) diagenetic reflux (Fs) from deposited sediments during their diagenesis and dissolution (Chaudhuri and Clauer, 1986; Goldstein and Jacobsen, 1987; Veizer, 1988; Palmer and Edmond, 1989). Secular variations of Sr isotopic composition in the global ocean show that the comparative redistribution of these fluxes was related to different source reservoirs, which influenced the general balance of matter entering marine basins in the past. This redistribution affected the bulk chemical composition of seawater and reflected essential changes in paleogeography and geodynamics. Secular variations of seawater composition are mainly dependant on the proportions between continental and mantle fluxes (Fc/Fm), which differ greatly in their Sr isotopic composition (Goldstein and Jacobsen, 1987; Veizer, 1988; Palmer and Edmond, 1989). According to available appraisals, the ranges of $^{87}Sr/^{86}Sr$ ratios are 0.7101 - 0.7119 for recent river runoff (Wadleigh *et al.*, 1985; Goldstein and Jacobsen, 1987; Palmer and Edmond, 1989) and 0.7031 - 0.7041 for modern hydrothermal solutions (Albarede *et al.*, 1981; Elderfield and Greaves, 1981; Michard *et al.*, 1984; Piepgrass and Wasserburg, 1985; Von Damm *et al.*, 1985). During different periods of geological history, the Sr isotopic ratios in the continental flux varied under the influence of tectonic environments and age or the compositional differences of provenances (Hodell *et al.*, 1990; Richter *et al.*, 1992; Berner and Rye, 1992); however, those in the mantle flux can only be lower, as is implied by the general evolutionary trend of Sr isotopic systems in the mantle. The diagenetic flux of strontium (Fs) from dissolving carbonates seems to be lower than 10% of its total input into oceans (Elderfield and Geiskes, 1982; Geiskes *et al.*, 1986), and its $^{87}Sr/^{86}Sr$ ratios are similar to those in seawater (Goldstein and Jacobsen, 1987; Kump, 1989; Palmer and Edmond, 1989; Richter and Liang, 1993). Accordingly, this flux is too weak to cause significant changes in seawater Sr isotopic composition, though it can serve as a buffer that somewhat limits their amplitudes.

Geological subjects applicable for studies of Sr isotopic variations in past marine basins should have low

Rb but high Sr concentrations, having formed under an Sr isotopic equilibrium with seawater. In addition, their Sr isotopic systems should be closed throughout the entire geological history. Carbonate shells of marine invertebrate as well as bio- and chemogenic carbonate rocks are the best Phanerozoic samples of this kind. Many of these samples have already been studied in order to plot the general variation curve of $^{87}Sr/^{86}Sr$ ratios of Phanerozoic basins, where they range between 0.7067 and 0.7089 (for review and references, see Faure, 1986).

The same information on Precambrian carbonate rocks (Veizer and Compston, 1976; Veizer *et al.*, 1983; Derry and Jacobsen, 1988; Derry *et al.*, 1989; Asmerom *et al.*, 1991; Pokrovskii and Vinogradov, 1991; Kaufman *et al.*, 1993) is more limited and not always confident, primarily because there are no carbonate skeletons in this part of the geological record. Moreover, the postdepositional changes of Sr isotopic composition are more probable for these rocks than for the Phanerozoic ones, and their ages are not constrained everywhere with the necessary precision. Consequently, secular variations of Sr isotopic composition in Precambrian oceans is known only in the general form of a curve ascending from the $^{87}Sr/^{86}Sr$ ratio range of 0.702 - 0.703 at the Archean–Proterozoic boundary to the value of 0.709 at the very end of the Proterozoic. The latter is preceded by a notable depression with a value range of 0.7054 - 0.7056, which corresponds to the Late Riphean (840 - 900 Ma) mantle events (Veizer *et al.*, 1983; Asmerom *et al.*, 1991). In contrast to this, very low $^{87}Sr/^{86}Sr$ ratios (0.7037 - 0.7052) have been reported in the Lower, Middle, and lower Upper Riphean dolomites of the Anabar massif (Pokrovskii and Vinogradov, 1991).

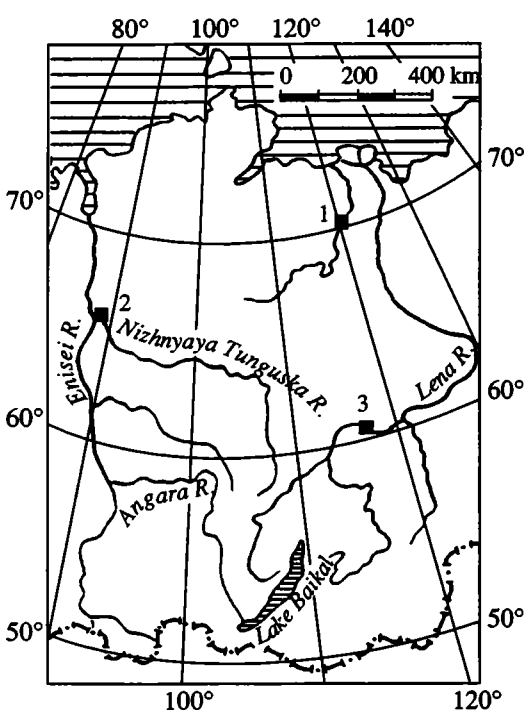

Fig. 1. Location scheme of studied regions. Uplifts: 1 – Olenek, 2 – Turukhansk, and 3 – Ura.

In this work, we present (1) an improved technique for estimating the initial Sr isotopic composition in carbonate rocks, which is based on a selective study of Sr fractions related to different rock phases with dissimilar isotopic compositions (Gorokhov *et al.*, 1992); (2) a geochemical and isotopic evaluation of the postdepositional alterations in the carbonate rock composition; (3) initial $^{87}Sr/^{86}Sr$ ratios for 51 limestone and dolomite samples from the Riphean, Vendian, and Lower Cambrian deposits of Siberia; and (4) a discussion of the significance of this data to obtain a better understanding of Sr isotopic system evolution in the Proterozoic oceans.

STRATIGRAPHIC POSITION OF STUDIED SAMPLES

The studied samples were collected in three Siberian regions: the Olenek, Turukhansk, and Ura uplifts (Fig. 1).

The Riphean deposits of the Olenek uplift form a very gently sloping (usually less than 3 - 5°) monocline with a series of low amplitude faults (Komar, 1966; Shpunt *et al.*, 1982; Shenfil' *et al.*, 1988). The deposits unconformably rest on pre-Riphean metamorphic rocks and granites and are transgressively overlain by subhorizontal Vendian strata, which bear Ediacarian fauna in the middle and the most ancient assemblages of small shelly fauna in the upper part of the succession. Upward the section, the Vendian deposits grade into sediments with fauna characteristic of the Tommotian Stage of the Lower Cambrian (Khomentovskii, 1985; Khomentovskii and Karlova, 1993). The color index of organic-walled microfossils of the Debengda Formation suggests that, in this area, the thermal history of the Riphean rocks was typical of that for platform regions. Samples for the Sr isotopic studies were collected in several localities: from a horizon 10 m above the base of the Lower Riphean Kyutingda Formation, the Kyutingda River headwaters (Sample 563/4a); from the base of the upper subformation of the Middle Riphean Arymas Formation (Sample 564/5); and from a lower part of the Debengda Formation (Sample 563/4), both times from the Khorbusuonka River localities.

In the Turukhansk region, the Riphean deposits are exposed along a marginal zone of the Siberian platform, where they form three blocks bounded by submeridional thrusts and transposed eastward. Inside the blocks, their beds, which are locally displaced by minor transverse faults, either dip gently westward in a monocline manner (usually less than 15 - 35°) or form an asymmetric synform with gentle eastern and vertical western limbs. The Vendian deposits unconformably

overlie various Riphean formations and grade upward the section into fossiliferous Lower Cambrian strata (Dragunov, 1963; Serebryakov, 1975). The color index of organic-walled microfossils from many rich assemblages shows that the Riphean deposits of the Turukhansk region were heated no more than 70 - 80°C. Carbonate samples for this study were collected from within two blocks with gently dipping beds exposed along the Nizhnyaya Tunguska and Bol'shaya Shorikha rivers. They represent all carbonate formations (Fig. 2) of the late Middle and early Late Riphean ages occurring in the region.

The Ura uplift comprises a series of brachyform folds prograding into the Siberian platform from an outer zone of the Patom fold belt, which frames the southern and southeastern margins of the platform (Salop, 1964). The folded rocks correspond to upper horizons of the Riphean. Near the marginal platform zones, they have more gentle inclinations and conformably underlie the Vendian–Cambrian strata of the platform sedimentary cover. The carbonate samples were collected from the Upper Riphean, Vendian, and Lower Cambrian deposits exposed near the Nokhtuisk settlement on the Lena River (Fig. 3).

It is difficult to precisely constrain the age of the studied samples, because isotopic datings of modern quality are scarce, and resolution of the biostratigraphic method is low when applied to the above intervals of geological records. These are the main obstacles for correctly positioning the Middle-Upper Riphean boundary in Siberian sections, which is still controversial (for references, see Komar, 1990; Shenfil', 1991).

Among all Riphean deposits sampled for this study, those of the Kyutingda, Arymas, and Debengda formations are the best correlative with the general Precambrian stratigraphic scale accepted in Russia (Semikhatov *et al.*, 1991). The Kyutingda Formation occurs at the base of the carbonate part of the region's Riphean succession and is separated from the pre-Riphean unconformity by clastic deposits of variable thicknesses (170 - 230 m in total) of the Osor-Khayata and Sygynakhtakh formations. The formation bears stromatolites characteristic of the Early Riphean age (Komar, 1966; Semikhatov and Serebryakov, 1983) and globular phyllosilicates (GPH), which occur in its middle part. Their K-Ar age, measured without mineralogical study of the analyzed material, is 1350 Ma (Kazakov *et al.*, 1965). On the basis of lithostratigraphic and paleontological criteria, the Kyutingda deposits are firmly correlated with similar sediments of the Anabar massif. In this region, the lower horizons of the succession, which correspond to the basal part of the Kyutingda Formation enclosing Sample 563/4a, contain glauconite dated by the Rb-Sr isochrone and K-Ar methods; the obtained age values are 1483 ± 5 and 1459 ± 10 Ma, respectively (Gorokhov *et al.*, 1991). Upward the section, the Kyutingda Formation is conformably overlain by siliciclastic-carbonate deposits of the Arymas (180 - 270 m) and Debengda (480 - 540 m) formations. In addition to the North Siberian endemics, both of them enclose stromatolites typical of the Middle Riphean (Komar, 1966; Shpunt *et al.*, 1982). The K-Ar ages, measured by Kazakov *et al.* (1965) using their GPH fractions, are 1165 and 1220 Ma for the Arymas Formation and 1040 and 1140 Ma for the Debengda Formation. However, recent age determinations using glauconites (including the Al-glauconite) from a lower part of the Debengda Formation (1265 Ma, the arithmetic mean for 8 mineral fractions, according to results of the K-Ar method and 1245 Ma according to the Rb-Sr isochrone measurement) indicate that this part of the section was formed approximately 1250 Ma ago (Baskakov *et al.*, 1991). Thus, our sample from the Kyutingda Formation (563/4a) is about 1480 Ma old, and the one from the Debengda Formation (563/4) is 1250 Ma old. The third sample from the Arymas Formation has an age that is intermediate between the second value and 1350 Ma, corresponding to the lower boundary age of the Middle Riphean (Semikhatov *et al.*, 1991), because all researchers of the Olenek Riphean rocks place this boundary at the base of the Arymas Formation.

In the Turukhansk region, all samples were collected from several levels of a thick (up to 3200 m), mostly carbonate succession (Fig. 2). The succession conformably rests on the siliciclastic Bezymyannyi Formation (up to 920 m thick), the base of which is unexposed. Being unconformably overlain by Vendian deposits, the succession is divided by an erosion surface (at the base of Derevnya Formation) into two sequences, which accumulated under different environments. The hiatus below the Derevnya Formation is marked by essential taxonomic changes in the stromatolite (Semikhatov and Serebryakov, 1983) and organic-walled microfossil assemblages (Veis, 1988; Yankauaskas *et al.*, 1989).

Until recently, only K-Ar datings of GPH from the Riphean rocks of the Turukhansk region were available.[1] The results obtained for various horizons of the succession are within the interval 800 - 910 Ma (for references, see Ivanovskaya *et al.*, 1973; Semikhatov and Serebryakov, 1983) and probably define only the minimum age limit of the host deposits. Using the Pb-Pb isochrone method, Ovchinnikova *et al.* (in press) recently determined an age value of 1017 ± 90 Ma for carbonate rocks of the Sukhaya Tunguska Formation (including some samples used in this study). To supplement this isotopic geochronology data, we will mention the K-Ar datings for GPH from correlative deposits occurring in the southeastern Enisei Ridge and Uchur-Maya region. The deposits are correlated on the basis of

[1] In addition, there are K-Ar data obtained by studying relatively coarse illite fractions and whole rock samples of claystones (Ivanovskaya *et al.*, 1973), which are not considered here, because later investigations (Bonhomme, 1987) have shown that such materials are useless for the age determination of sedimentation processes.

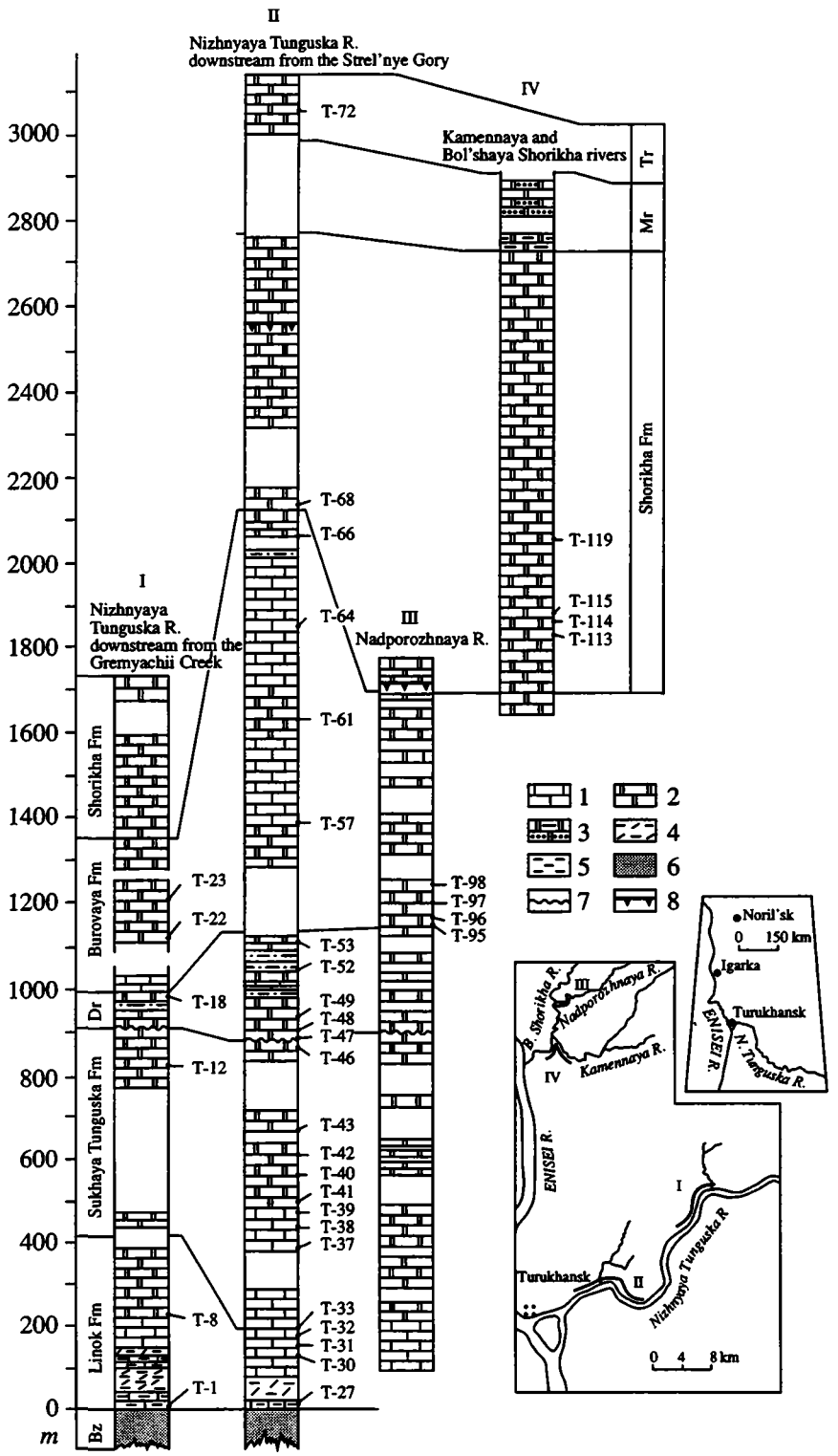

Fig. 2. Lithostratigraphic successions of the Riphean deposits in the Turukhansk uplift. Positions of the studied samples are indicated. 1 – limestones; 2 – dolomites; 3 – clayey and sandy limestones; 4 – alternating shales, limy shales, and limestones; 5 – shales; 6 – shales and sandstones; 7 – regional unconformity; 8 – faults. Formations: Bz – Bezymyannyi; Dr – Derevnya; Mr – Miroedikha; Tr – Turukhansk. Numbers I - IV in the map show the locations of the studied sections.

a surprising coincidence in their superposition, the similar succession of stromatolite assemblages, and characteristic organic-walled microfossils that appear at certain levels of the sections under comparison (for review and references, see Semikhatov and Serebryakov, 1983; Shenfil', 1991; Semikhatov, 1991). For our present study, we note the following: (1) the K-Ar ages of units correlative with the Bezymyannyi Formation (Kazakov and Knorre, 1973; Semikhatov and Serebryakov, 1989) are either 1100 Ma (the Pogoryui Formation of the Enisei Ridge) or 1000 - 1160 Ma (the Totta Formation in the Kerpyl' Group of the Uchur-Maya region); (2) the boundary between the Kerpyl' and Lakhanda groups of the Uchur-Maya region corresponds to the lower boundary of the Derevnya Formation in the Turukhansk uplift and is approximately 1000 Ma old (for details, see Semikhatov and Serebryakov, 1983, pp. 163 - 164; Shenfil', 1991); and (3) K-Ar age values of various horizons of the Lakhanda Group range between 850 and 950 Ma, while in the Enisei Ridge, the beds correlative to the Derevnya Formation are 900 - 940 Ma old (Lebedev *et al.*, 1980).

The position of the Middle-Upper Riphean boundary in the correlated Siberian sections is controversial. Some researchers place it at the base of the Kerpyl' Group of the Uchur-Maya region, thus assigning the entire pre-Vendian succession of the Turukhansk uplift to the Upper Riphean (Khomentovskii *et al.*, 1985; Veis, 1988; Komar, 1990; Shenfil', 1991). This opinion is based on criteria of historical geology and/or some paleontological data, such as the occurrence of new morphotype and dimension classes of organic-walled microfossils in the Kerpyl' Group and its counterparts, where they are accompanied by the appearance of some stromatolite forms characteristic of lower horizons of the Upper Riphean stratotype. However, this viewpoint implies a revision of the Uralian model of the Middle-Upper Riphean boundary, which has been accepted in the standard scale of the Precambrian of North Eurasia within the limits of the former USSR (Semikhatov *et al.*, 1991). In addition, this would increase the boundary age from 1000 to 1200 Ma. It must be noted that many of the above innovations in microfossil assemblages are also known from the Ust'-Il'ya and Kotuikan formations of the Anabar massif (Veis and Vorob'eva, 1992), which are Early Riphean in age according to the composition of the stromatolites and available Rb-Sr and K-Ar datings of glauconite (Komar, 1966; Shpunt *et al.*, 1982; Gorokhov *et al.*, 1991).

The second viewpoint, which is accepted in this work, suggests that the Middle-Upper Riphean boundary be placed at the base of the Derevnya Formation of the Turukhansk region and correlative units (*Resheniya ...*, 1983; Semikhatov and Serebryakov, 1983; Semikhatov, 1991). This opinion is based on two groups of paleontological remains appearing in the Derevnya Formation and Lakhanda Group: acanthomorphic acritarchs of the *Trachyhystrichosphaera* genus, which are the markers (Sergeev, 1992; Knoll *et al.*, 1991) of the Upper Riphean lower boundary (= the Neoproterozoic lower boundary) and some stromatolite taxa characteristic of lower horizons of the Upper Riphean stratotype (the Karatau Group of the southern Urals). This viewpoint is also supported by the current estimation of the isotopic age of the lower boundary of the Derevnya, Lakhanda, and Karatau successions (Semikhatov, 1991).

Thus, the studied samples from the Linok and Sukhaya Tunguska formations belong to the upper horizons of the Middle Riphean, with an age of 1000 - 1100 Ma. Those from the Derevnya, Burovaya, Shorikha, and Turukhansk formations correspond to the lower Upper Riphean, i.e., they are younger than 1000 Ma but older than 850 - 800 Ma.

A sample series collected within the Ura uplift represents deposits that are less ancient. The Tolbachan and uppermost Macha deposits of the region contain fauna of the Atdabanian Stage of the Lower Cambrian (*Yarusnoe ...*, 1984; Shenfil', 1991). The underlying Tinnaya Formation should be assigned to the middle part of the Vendian, according to its stratigraphic position and identified stromatolites and microfossils (Dol'nik *et al.*, 1990; Shenfil', 1991; Stanevich and Faizulina, 1992). In this region, all stratigraphers place the lower Vendian boundary at the base of the Dzherba Formation, which conformably underlies the Tinnaya Formation and grades downward into the Zhuya Group (Fig. 3). Accordingly, this group comprises the upper horizons of the Riphean. It is difficult to constrain its age more precisely, because no isotopic age data is available for the Riphean deposits of the Baikal–Patom region. Moreover, they contain abundant but endemic stromatolites and microfossil assemblages with very poor taxonomic diversity. Thus, the deposit's age is still under discussion.

Some geologists traditionally believe that the Zhuya Group comprises the entire Upper Riphean and, consequently, refer the underlying Dal'nyaya Taiga Group to the upper part of the Middle Riphean (Dol'nik, 1982; Dol'nik *et al.*, 1990). Another way of approaching the problem is to consider the latter group as a unit of the lower Upper Riphean, as it contains some representative microfossils (Stanevich and Faizulina, 1992). The third opinion that correlates the Dal'nyaya Taiga and Zhuya groups with the pre-Vendian deposits of the Enisei Ridge, which discordantly overlay granitoids with an isotopic age of about 850 Ma, is based on a long chain of lithostratigraphic correlations (Khomentovskii, 1976; Khomentovskii *et al.*, 1985; Shenfil', 1991). Still, our analysis of the available information concluded that the Zhuya Group is a unit of the upper Upper Riphean and probably is younger than 800 Ma.

Thus, one can see that the available data on isotopic geochronology and paleontology constrains only broad age intervals for most of the sampled units. Considering our limited knowledge of the age of the upper Precambrian of Siberia, we must choose one or another interpretation procedures to estimate more precise

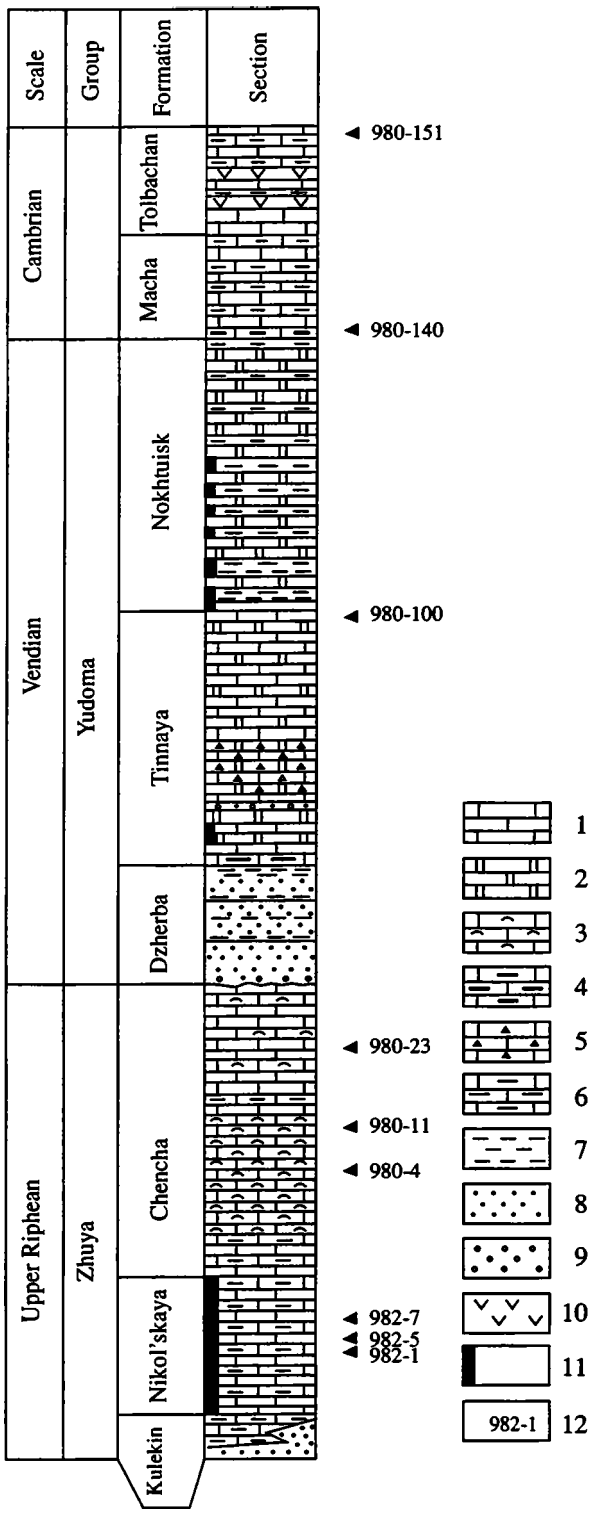

Fig. 3. Lithostratigraphic successions of the upper Upper Riphean, Vendian, and Lower Cambrian deposits in the Ura uplift. Positions of the studied samples are indicated.

1 – limestones; 2 – dolomites; 3 – stromatolite limestones; 4 – bituminous limestones; 5 – brecciated limestones; 6 – clayey limestones; 7 – shales; 8 – shales and sandstones; 9 – gravelstones; 10 – diabase sills; 11 – red deposits; 12 – sample no.

confinements of particilar samples to the geochronometric (numerical) scale. The simplest way is to assume that deposit thicknesses are proportional to the time intervals of their accumulation. Of course, this is a very rough approximation of the real relations, especially when a succession consists of alternating units with different lithologic characteristics. Nevertheless, without sedimentological data on relative sedimentation rates, this method seems more acceptable than any other speculative corrections. It was used to estimate the age of samples from the Kyutingda and Arymas formations of the Olenek uplift and from the Zhuya Group and the Tinnaya Formation of the Ura uplift.

The situation involving samples from the Turukhansk uplift is somewhat easier. Recent sedimentological data (Petrov, 1993a, 1993b) suggest that the succession of the Bezymyannyi, Linok, and Sukhaya Tunguska formations was formed during a large regressive cycle comprising several stages of basin filling during a gradual decrease in the rate of sedimentation. This conclusion and the inferred paleotectonic setting of the region permitted the age values of samples from the Linok and Sukhaya Tunguska formations to be constrained using an empirical equation relating the deposit thickness and sedimentation rate, which has been inferred for the North American passive margin (Sleep, 1971) and already used in some studies of Sr isotopic composition in Precambrian carbonates (Derry *et al.*, 1989; Asmerom *et al.*, 1991; Kaufman *et al.*, 1993). The younger Riphean strata (above the Sukhaya Tunguska Formation) of the Turukhansk region were mainly deposited on an actively subsiding carbonate platform. The only exception is the Miroedikha Formation, which is related to a short period of basin deepening with a higher input of siliciclastic material. Disregarding this episode, we accept that in the Turukhansk region all Riphean units above the Sukhaya Tunguska Formation were deposited under fairly constant sedimentation rates. If the analyzed samples were collected from two or three separate sections of a formation, we correlated their levels with the most complete stratigraphic section of the formation using litho- and chemostratigraphic criteria. Such reference sections served to constrain all sample ages by means of the interpolation procedure described above. Of course, the feasibility of the procedure causes the resulting geochronometric values to be arbitrary.

In this study, we accept the placement of the lower Cambrian boundary at the base of the Tommotian Stage, which is traditional for Russian scientists. In the Ura anticlinorium, this boundary position was found at the base of the Macha Formation (Dol'nik, 1982; Dol'nik *et al.*, 1990; Stanevich and Faizulina, 1992). Recent isotopic measurements well defined the age of the boundary to be 535 Ma (for review and original data, see Bowring *et al.*, 1993), while the boundary between the Tommotian and Atdabanian stages was dated at 530 Ma. The estimated ages of the studied samples are presented in Table 1. Actually, these ages

Table 1. Stratigraphic positions and estimated ages of studied samples

Sample no.	Rock*	Estimated age, Ma	Sample no.	Rock*	Estimated age, Ma	Sample no.	Rock*	Estimated age, Ma
Tolbachan Formation			Burovaya Formation			Sukhaya Tunguska Formation		
980-151	L	530	T-66	L	922	T-46	D	1004
Macha Formation			T-64	L	935	T-12	D	1008
980-140	L	540	T-74	D	935	T-43	D	1048
Tinnaya Formation			T-73	L	938	T-42	D	1055
980-100	L	590	T-61	D	950	T-40	D	1061
Chencha Formation			T-23	D	956	T-41	D	1068
980-23	L	669	T-22	D	962	T-39	L	1069
980-11	L	697	T-57	D	968	T-38	L	1072
980-4	L	711	T-98	D	978	T-37	L	1077
Nikol'skaya Formation			T-97	D	981	Linok Formation		
982-7	L	761	T-96	L	983	T-33	L	1091
982-5	L	765	T-95	L	984	T-32	L	1091
982-1	L	767				T-31	L	1092
Turukhansk Formation			Derevnya Formation			T-8	L	1094
T-72	D	856	T-53	D	984	T-30	L	1094
Shorikha Formation			T-18	D	986	T-27	D	1096
T-119	D	903	T-52	L	989	T-1	D	1100
T-115	D	910	T-49	D	996	Debengda Formation		
T-114	D	911	T-48	D	998	563/4	L	1250
T-113	D	912	T-47	D	999	Arymas Formation		
						564/5	L	1300
						Kyutingda Formation		
						563/4a	L	1480

* L - limestone; D - dolomite.

should not be considered at the precision level of the shown numeric values, but the differences between these values seem to fairly well illustrate the time intervals separating the studied samples.

ANALYTICAL PROCEDURE

Carbonate rock samples were first studied by using the X-ray diffractomemter DRON-UM-1 with a Co-emitter to determine the proportions between their calcite and dolomite fractions on the basis of peak area (112) measurements (3.03 Å for calcite, and 2.88 Å for dolomite). The samples dominated by one of these minerals were then selected for further analysis. The only exception was Sample T-73, which is composed of a calcite-dolomite mixture. After dissolving the samples in HCl, chemical weight analysis was used to measure Ca and Mg contents, while Mn and Fe concentrations were determined by the atomic-absorbtion method. Concentrations of the major and trace elements characterizing the carbonate fractions of the studied samples are shown in Table 2. To estimate them, we corrected the measured concentrations by accounting for the nonsoluble silicate fractions of the samples. The last column of the table includes data on oxygen isotopic composition obtained at Harvard University (Knoll *et al.*, in press) and at the Institute of Precambrian Geology and Geochronology, Russian Academy of Sciences (unpublished data kindly submitted by Vinogradov and Podkovyrov).

By studying the Rb-Sr isotopic systematics, the powdered rock samples were preliminarily treated with an ammonium acetate (NH_4OAc) solution (Gorokhov *et al.*, 1992). It has been suggested (Walls *et al.*, 1977; Gao, 1990) that such a treatment removes readily soluble chemical elements situated outside the crystal lattice of primary carbonate minerals. Each powdered sample (about 0.1 g), placed in a Teflon test tube with 3 ml of 1 N solution of NH_4OAc, was subjected to ultrasonic vibration for 10 minutes. The resulting solution and nonsoluble phase were separated by centrifuging at 8000 rpm for 30 minutes. After a dual operation of this kind, the coupled centrifugates were used in further analysis. To measure the Rb and Sr concentrations, as

Table 2. Geochemical characteristics of carbonate fractions in studied samples of limestones and dolomites

Sample no.	Mg/Ca	Mn/Sr*	Fe/Sr*	Ca/Sr*	$\delta^{18}O$, ‰ PDB**
1	2	3	4	5	6
Tolbachan Formation					
980/151	0.003	0.72	3.7	2350	–
Macha Formation					
980/140	0.025	1.93	13.3	2040	−8.9
Tinnaya Formation					
980/100	0.052	1.31	22.3	2210	−19.7
Chencha Formation					
980/23	0.02	0.61	2.7	911	–
980/11	0.024	0.04	1.3	229	–
980/4	0.038	0.08	1.9	292	–
Nikol'skaya Formation					
982/7	0.019	1.32	9.8	1130	–
982/5	0.028	1.37	2.0	1030	–
982/1	0.031	2.18	2.9	1150	–
Turukhansk Formation					
T-72	0.597	1.57	12.8	5260	–
Shorikha Formation					
T-119	0.635	4.34	59.2	8100	−7.0
T-115	0.597	8.05	91.2	11420	−6.7
T-114	0.595	6.85	34.2	4100	−6.2
T-113	0.622	3.59	35.9	7720	−7.4
Burovaya Formation					
T-66	0.022	1.79	8.6	2970	−8.2
T-64	0.005	0.02	0.2	371	−6.4
T-74	0.589	10.46	79.8	5940	−4.9
T-73	0.108	0.33	6.0	1400	−7.3
T-61	0.616	8.05	95.2	15320	−10.2
T-23	0.577	1.88	10.8	3800	−6.4
T-22	0.585	1.34	17.6	3980	−6.0
T-57	0.61	3.29	164.5	13640	−11.3
T-98	0.592	5.7	88.7	13740	−8.2
T-97	0.579	8.48	152.7	8520	−8.7
T-96	0.053	0.11	3.0	479	−6.2
T-95	0.052	0.54	12.6	3260	−5.5
Derevnya Formation					
T-53	0.572	6.17	65.3	8000	−8.6
T-18	0.584	16.98	66.6	4790	−8.5
T-52	0.023	4.79	27.7	4660	−9.0
T-49	0.572	6.34	70.4	5140	−7.3
T-48	0.511	19.69	187.0	3830	−5.7
T-47	0.603	4.22	29.3	4050	−6.2
Sukhaya Tunguska Formation					
T-46	0.587	3.0	42.8	4680	−5.1
T-12	0.608	0.97	10.8	3060	−5.3

Table 2. (Contd.)

1	2	3	4	5	6
T-43	0.559	0.59	9.5	2640	−3.5
T-42	0.598	0.66	10.4	2890	−5.4
T-40	0.556	1.49	19.4	5300	−5.9
T-41	0.598	1.76	21.7	11000	−6.7
T-39	0.013	1.1	32.3	10640	−9.2
T-38	0.009	0.05	0.7	993	−9.8
T-37	0.029	0.04	0.8	723	−8.9
		Linok Formation			
T-33	0.05	0.53	16.6	2720	−8.1
T-32	0.003	2.58	52.6	3000	−8.7
T-31	0.01	0.7	14.4	2140	−7.8
T-8	0.006	1.19	37.9	1250	−9.3
T-30	0.003	0.64	14.1	2380	−8.3
T-27	0.344	82.59	565.7	2130	−9.9
T-1	0.415	157.49	1257.0	3560	–
		Debengda Formation			
563/4		11.88	20.0		–
		Arymas Formation			
564/5		3.1	52.4		–
		Kyutingda Formation			
563/4a		6.55	15.9		–

Notes: * Sr concentrations are calculated as weighted sum of the "readily and less soluble" Sr (see Table 3 for original data).

** Data after Knoll *et al.* (in press); unpublished data for samples 980/100 and 980/140 were submitted byD.P. Vinogradov and V.N. Podkovyrov.

well as the Sr isotopic composition in the insoluble residues, we treated them with a 5% solution of acetic acid (HOAc), which was added by drops up to a volume of 3 ml. To dissolve the main part of the carbonate phases, the treatment was as long as 30 minutes for the limestones and 24 hours for the dolomites. Further operations of ultrasonic and centrifugal machining were similar to those described above.

The aliquots of centrifugates, obtained after both the NH_4OAc and HOAc treatments, were used to analyze the Rb and Sr concentrations by the isotope dilution technique. Centrifugate portions that remained after aliquote removal were evaporated until they were dry and then transformed into chloride solution by boiling with concentrated HCl. The standard ion-exchange technique with cation exchange resin Dowex AG50x8 was used to extract the Sr for isotopic analysis.

The weight of the insoluble residue after the original sample was treated with the NH_4OAc and HOAc solutions was usually below 10%, except for several samples (Table 3) with higher values. As implied by the X-ray diffractometry data, the insoluble residues consist mainly of quartz, sometimes accompanied by accessory feldspar and dolomite.

A single-collector mass spectrometer MI-1320 was used to measure the isotopic composition of Rb and Sr for an evaluation of their concentrations. The Sr isotopic composition was analyzed with the multicollector mass spectrometer Finnigan MAT-261 under a regime of simultaneous measurements of all isotope ion currents. During the employment period of this mass spectrometer, the mean value of the $^{87}Sr/^{86}Sr$ ratio, normalized to the $^{86}Sr/^{88}Sr = 0.1194$ of the standard Sr carbonate SRM-987 of the USA National Bureau of Standards, was 0.71026 ± 0.00001 (2σ m for 23 analyses). For analyzed samples, the within-run uncertainty of the $^{87}Sr/^{86}Sr$ ratio measurement during analysis was below 0.00007 and corresponded on average to 0.00004 (2σ).

GEOCHEMICAL AND ISOTOPIC EVALUATION OF POSTDEPOSITIONAL ALTERATIONS IN CARBONATE ROCKS

Postdepositional alterations of carbonate minerals of marine origin may sometimes result in essential changes in their Sr isotopic composition. In the studied samples lacking features of hydrothermal or metamorphic alterations, such changes could be caused primarily by late diagenetic (catagenetic) processes.

Table 3. Rb-Sr data for soluble fractions of limestones and dolomites

Sample no.	Dominant mineral*	Chemical reagent	Percentage of dissolved fraction, %	Rb, ppm	Sr, ppm	$^{87}Sr/^{86}Sr$	$^{87}Sr/^{86}Sr$ (measured)	$^{87}Sr/^{86}Sr$ (initial)
1	2	3	4	5	6	7	8	9
			Tolbachan Formation					
980-151	C	NH_4OAc	8	3.09	137	0.0651	0.70865	0.70816
		HOAc	88	0.23	171	0.0038	0.70857	0.70854
			Macha Formation					
980-140	C	NH_4OAc	6	11.3	246	0.1334	0.70920	0.70818
		HOAc	87	0.27	184	0.0042	0.70858	0.70855
			Tinnaya Formation					
980-100	C	NH_4OAc	10	7.92	–	–	0.70984	–
		HOAc	75	0.47	203	0.0067	0.70859	0.70853
	C	NH_4OAc	12	5.86	143	0.1183	0.70949	0.70850
		HOAc	81	0.45	167	0.0077	0.70855	0.70849
			Chencha Formation					
980-23	C	NH_4OAc	7	1.64	495	0.0096	0.70879	0.70870
		HOAc	91	0.28	417	0.0019	0.70855	0.70853
980-11	C	NH_4OAc	10	0.67	1028	0.0019	0.70833	0.70831
		HOAc	88	0.42	1744	0.0007	0.70826	0.70825
980-4	C	NH_4OAc	7	1.20	1010	0.0034	0.70824	0.70821
		HOAc	91	0.22	1311	0.0005	0.70812	0.70811
			Nikol'skaya Formation					
982-7	C	NH_4OAc	8	2.19	374	0.0169	0.70922	0.70904
		HOAc	86	0.12	340	0.0010	0.70832	0.70831
	C	NH_4OAc	9	2.15	278	0.0224	0.70903	0.70879
		HOAc	84	0.13	339	0.0011	0.70837	0.70836
982-5	C	NH_4OAc	10	1.59	580	0.0080	0.71066	0.71057
		HOAc	87	0.14	363	0.0011	0.70932	0.70931
982-1	C	NH_4OAc	8	0.83	348	0.0068	0.71055	0.71048
		HOAc	90	0.24	330	0.0021	0.70942	0.70940
			Turukhansk Formation					
T-72	D	NH_4OAc	3	1.23	84.2	0.0423	–	–
		HOAc	97	0.03	41.2	0.0023	0.70625	0.70622
			Shorikha Formation					
T-119	D	NH_4OAc	2	5.64	83.7	0.1867	0.70798	0.70559
		HOAc	86	0.24	24.2	0.0281	0.70630	0.70594
T-115	D	NH_4OAc	3	2.77	70.2	0.1139	0.70787	0.70640
		HOAc	78	0.09	17.4	0.0157	0.70636	0.70616
T-114	D	NH_4OAc	4	1.67	83.8	0.0585	0.70726	0.70650
		HOAc	96	0.06	52.1	0.0034	0.70574	0.70570
T-113	D	NH_4OAc	4	1.24	59.0	0.0614	0.70704	0.70624
		HOAc	96	0.03	26.9	0.0031	0.70580	0.70576
			Burovaya Formation					
T-66	C	NH_4OAc	5	2.20	189	0.0338	0.70734	0.70690
		HOAc	93	0.14	128	0.0031	0.70603	0.70599
T-64	C	NH_4OAc	9	0.38	808	0.0014	0.70527	0.70525
		HOAc	91	0.13	1105	0.0003	0.70523	0.70523

Table 3. (Contd.)

1	2	3	4	5	6	7	8	9
T-74	D	NH_4OAc	2	4.36	110	0.1145	0.71092	0.70940
		HOAc	95	0.06	35.1	0.0050	0.70634	0.70627
T-73	C	NH_4OAc	6	3.15	658	0.0146	0.70653	0.70634
		HOAc	93	0.05	219	0.0006	0.70631	0.70630
T-61	D	NH_4OAc	2	5.62	65.0	0.2500	0.71092	0.70755
		HOAc	95	0.10	12.6	0.0230	0.70624	0.70593
T-23	D	NH_4OAc	2	2.83	206	0.0402	0.70730	0.70675
		HOAc	98	0.03	56.4	0.0017	0.70627	0.70625
T-22	D	NH_4OAc	2	4.94	164	0.0884	0.70751	0.70630
		HOAc	98	0.03	52.4	0.0019	0.70609	0.70606
T-57	D	NH_4OAc	3	7.37	53.8	0.3967	0.71139	0.70594
		HOAc	56	0.19	14.2	0.0385	0.70752	0.70699
T-98	D	NH_4OAc	3	1.87	33.8	0.1600	0.70848	0.70626
		HOAc	71	0.05	15.1	0.0102	0.70625	0.70611
T-97	D	NH_4OAc	3	8.77	43.7	0.5799	0.71473	0.70665
		HOAc	57	0.30	24.7	0.0347	0.70647	0.70599
T-96	C	NH_4OAc	5	3.24	103	0.0913	0.70773	0.70646
		HOAc	92	0.07	818	0.0024	0.70572	0.70569
T-95	C	NH_4OAc	6	7.03	126	0.1615	0.70754	0.70528
		HOAc	90	0.16	114	0.0041	0.70543	0.70537
			Derevnya Formation					
T-53	D	NH_4OAc	3	6.14	73.9	0.2411	0.71085	0.70748
		HOAc	96	0.06	27.1	0.0063	0.70610	0.70601
T-18	D	NH_4OAc	3	17.4	586	0.0869	0.71008	0.70886
		HOAc	95	0.16	28.1	0.0452	0.70825	0.70762
T-52	C	NH_4OAc	5	1.92	23.1	0.2405	0.70849	0.70511
		HOAc	89	0.25	86.3	0.0084	0.70625	0.70613
T-49	D	NH_4OAc	4	0.94	58.7	0.0461	0.70789	0.70724
		HOAc	94	0.06	42.1	0.0044	0.70590	0.70584
T-48	D	NH_4OAc	9	10.7	89.1	0.3463	0.71225	0.70734
		HOAc	72	0.86	54.1	0.0459	0.70723	0.70658
T-47	D	NH_4OAc	4	1.00	51.0	0.0570	0.70740	0.70659
		HOAc	95	0.03	55.2	0.0014	0.70594	0.70592
			Sukhaya Tunguska Formation					
T-46	D	NH_4OAc	3	2.45	67.2	0.1053	0.70660	0.70510
		HOAc	82	0.07	46.5	0.0041	0.70571	0.70565
T-12	D	NH_4OAc	4	1.42	98.0	0.042	0.70682	0.70622
		HOAc	93	0.06	69.0	0.0025	0.70588	0.70584
T-43	D	NH_4OAc	3	1.20	138	0.0249	0.70632	0.70595
		HOAc	97	0.04	84.0	0.0013	0.70584	0.70582
T-42	D	NH_4OAc	8	7.75	81.6	0.2750	0.70613	0.70201
		HOAc	85	0.83	75.1	0.0320	0.70580	0.70532
T-40	D	NH_4OAc	6	0.92	72.0	0.0369	0.70722	0.70666
		HOAc	89	0.06	41.1	0.0041	0.70612	0.70606
T-41	D	NH_4OAc	9	0.16	16.4	0.0272	0.70757	0.70716

Table 3. (Contd.)

1	2	3	4	5	6	7	8	9
		HOAc	89	0.04	20.3	0.0053	0.70587	0.70579
T-39	C	NH_4OAc	6	0.20	43.0	0.0138	0.70645	0.70624
		HOAc	87	0.02	36.2	0.0016	0.70583	0.70581
T-38	C	NH_4OAc	37	0.07	41.8	0.0047	0.70652	0.70645
		HOAc	61	0.05	612	0.0002	0.70585	0.70585
T-37	C	NH_4OAc	5	0.78	583	0.0039	0.70595	0.70589
		HOAc	94	0.16	530	0.0009	0.70568	0.70567
			Linok Formation					
T-33	C	NH_4OAc	6	1.90	142	0.0386	0.70797	0.70737
		HOAc	88	0.23	139	0.0048	0.70659	0.70652
T-32	C	NH_4OAc	6	6.12	160	0.1108	0.71074	0.70902
		HOAc	75	0.32	133	0.0069	0.70708	0.70697
	C	NH_4OAc	8	4.26	107	0.1154	0.71033	0.70854
		HOAc	71	0.22	132	0.0047	0.70701	0.70694
T-31	C	NH_4OAc	8	1.84	124	0.0427	0.70756	0.70690
		HOAc	85	0.14	192	0.0021	0.70614	0.70611
T-8	C	NH_4OAc	30	4.62	33.6	0.3981	0.70999	0.70381
		HOAc	48	1.58	468	0.0098	0.70636	0.70621
	C	NH_4OAc	11	12.4	184	0.1960	0.70889	0.70585
		HOAc	67	0.90	314	0.0083	0.70625	0.70612
T-30	C	NH_4OAc	8	3.32	130	0.0738	0.70747	0.70633
		HOAc	86	0.21	172	0.0036	0.70607	0.70601
T-27	FD	NH_4OAc	5	31.6	215	0.4263	0.72305	0.71642
		HOAc	64	1.24	102	0.0353	0.71664	0.71609
T-1	FD	NH_4OAc	4	60.7	203	0.8686	0.72284	0.70927
		HOAc	57	2.84	46.4	0.1772	0.71451	0.71174
			Debengda Formation					
563/4	C	NH_4OAc	4	7.03	216	0.0944	0.70934	0.70766
		HOAc	90	0.29	157	0.0053	0.70499	0.70490
			Arymas Formation					
564/5	C	NH_4OAc	4	5.20	157	0.0963	0.71047	0.70869
		HOAc	94	0.57	144	0.0115	0.70634	0.70613
			Kyutingda Formation					
563/4a	C	NH_4OAc	4	4.95	135	0.1062	0.71381	0.71158
		HOAc	96	0.11	38.1	0.0081	0.70482	0.70465

* C – calcite; D – dolomite; FD – Fe-dolomite, or calcite + ankerite.

The geochemical approach to evaluating the degree of postdepositional alterations in carbonate rocks is the most promising (Asmerom *et al.*, 1991; Derry *et al.*, 1992) due to relevant changes in the concentrations of many trace elements. Brand and Veizer (1981) have shown that the Mn/Sr ratio is a very sensitive indicator of secondary alterations, because carbonates become enriched in Mn and Fe but depleted in Sr during late freshwater diagenesis (catagenesis), which is usually responsible for the isotopic system disturbances in these minerals. The enrichment and depletion degree is dependent on the element distribution coefficients and the differences in the Ca/Mn, Ca/Fe, and Ca/Sr ratios of seawater and diagenetic fluid. Below, using geochemical and isotopic data, we evaluate the extent of postdepositional changes in the studied limestones and dolomites.

Limestones. The Ca/Sr and Mn/Sr ratios in the studied limestones (Table 2) have wide variation ranges. Half of the samples revealed values higher than the limits of ≤1000 and ≤1, respectively, which have been suggested as typical for the unaltered Upper Riphean carbonates from Spitzbergen and eastern Greenland (Derry *et al.*, 1992). However, somewhat higher limits

(≤3200 for Ca/Sr and ≤1.7 for Mn/Sr) have been considered for the unaltered samples of the Upper Riphean limestones of northern Canada (Asmerom *et al.*, 1991). If the latter values are correct, then the limestones under analysis should be regarded as mostly unaffected by significant secondary alterations, except those with very high Mn/Sr and high Fe/Sr ratios, represented by samples (563/4, 563/4a, and 564/5) from the Kyutingda, Arymas, and Debengda formations of the Olenek uplift and the Derevnya Formation of the Turukhansk uplift (Sample T-52). Remarkably elevated Fe/Sr ratios were also calculated for limestones from the Linok Formation, despite their low Mn/Sr values. Data on oxygen isotope composition obtained for samples from the Turukhansk uplift (Knoll *et al.*, in press) suggest that they are mostly unaltered, except the samples from the Linok Formation, which revealed a positive correlation between $\delta^{18}O$ and $\delta^{13}C$ parameters, indicating that the carbon isotopic composition was changed in samples with the lowest $\delta^{18}O$ values (T-1, T-27, and T-32).

Additional information on secondary alterations of limestones is inferable from the Rb-Sr systematics for sample phases soluble in NH_4OAc and HOAc.

The data of Table 3 shows that limestone fractions soluble in NH_4OAc are usually below 10%. Higher proportions up to 30 - 37% (samples T-8 and T-38) represent an artifact, because the decantation procedure was used instead of centrifuging to separate the solutions and insoluble residues of these samples. As a result, a part of the solid residue entered the acetate solutions in the suspended state, thus causing the apparent increase of dissolved fraction. Therefore, we may conclude that NH_4OAc dissolves only a minor part of limestone carbonates; the rest goes into solution after the next treatment with HOAc. These results support the earlier conclusion (Schroeder, 1969; Walls *et al.*, 1977) that carbonate rocks contain mineral constituents with different trace element concentrations and variable degrees of solubility in chemical reagents.

In all the studied cases, sample fractions dissolved in NH_4OAc differed greatly in their Rb-Sr systematics from those dissolved in HOAc. Consequently, the limestones consist of two carbonate phases: (1) the readily soluble one, going into NH_4OAc solution; and (2) the more resistant phase, soluble in HOAc and stronger acids. The Rb concentrations and Rb/Sr ratios are much (sometimes ten-fold) higher in the phase soluble in NH_4OAc than in the fraction of lesser solubility (Figs. 4a and 4c). In addition, the $^{87}Sr/^{86}Sr$ ratio is always higher in the readily soluble phase than in the major sample part (Fig. 5a). Therefore, the two-stage treatment of samples with ammonium acetate (NH_4OAc) and acetic acid (HOAc) extracts two different Sr systems.

The readily soluble Sr can occur in silicate and carbonate rock constituents as ions: (1) adsorbed on crystal surfaces and in supercapillary interstices, (2) dissolved in fluid inclusions, and (3) incorporated into superficial and inner mineral inclusions. Preliminary treatment with 0.2 N solution of NH_4OAc has also been used to remove the readily soluble Sr, which was related to clay components of the Upper Cambrian and Lower Ordovician dolomites from Oklahoma and Texas, USA (Gao, 1990; Kupecz and Land, 1991). Actually, experimental treatment of the Lower Cambrian blue clays from northern Estonia with 1 N solution of NH_4OAc (Gorokhov *et al.*, 1989) has shown that this procedure is able to remove more than half of the available Sr leached from clay mineral surfaces, the expandable layers of these minerals, and, probably, their interlayer levels. However, the carbonate rocks investigated in the present study are free of clay minerals, as follows from the X-ray diffractometry of their insoluble residues. As to whether quartz and feldspar compose the insoluble fractions of limestones, their structure and chemical composition rule out any significant Sr leaching by the treatment of 1 N solution of NH_4OAc under pH 7.0. For example, after leaching under these conditions, only 0.9% Rb and 2.1% Sr went into acetate solution from the feldspar standard 70a (NBS, USA), with concentrations of 521 ppm Rb and 65.5 ppm Sr. We must also note that the analyzed limestone samples generally contain 0.5 - 4.4% of insoluble residue; three samples yielded 6.9 - 9.3% of this fraction, and only in two limestones (T-32, T-8) was its proportion as high as 20.7 and 21.4%. In addition, the feldspar was always a minor component of the silicate admixture. Bearing this in mind, we are sure that the carbonate components of limestones represent the main (if not the only) source of readily soluble Sr (and Rb). The data on samples T-64 and 563/4a, which entirely lack an insoluble silicate fraction, favors this conclusion well.

As can be seen from Table 3 and Fig. 4b, Sr concentrations in the phase soluble in NH_4OAc are of the same or, usually, higher levels as compared to those in the less soluble limestone fractions. This suggests that readily soluble Sr is mostly a component of fluid or mineral inclusions in the carbonate material, despite some possible admixture of ions adsorbed on the surface of mineral grains.

The Sr^{2+} concentrations in some natural brines can be as high as 1000 - 2000 ppm (McNutt *et al.*, 1987, 1990; Russell *et al.*, 1988). However, interstitial waters of deep-sea carbonate sediments usually contain less than 100 ppm Sr^{2+} (Sayles and Manheim, 1975; Hawkesworth and Elderfield, 1978; Baker *et al.*, 1982; Elderfield *et al.*, 1982; Geiskes *et al.*, 1986, 1990). One should expect similar Sr^{2+} abundances in fluid inclusions entraped by the Proterozoic carbonate rocks used in this study. Nevertheless, the data in Table 3 shows that easily soluble limestone phases often have notably higher Sr concentrations. This suggests that many of the studied samples contain an easily soluble component enriched in strontium.

This component with an Sr concentration comparable to that in the less soluble phase seems to be a product of the reaction between this phase and circulating

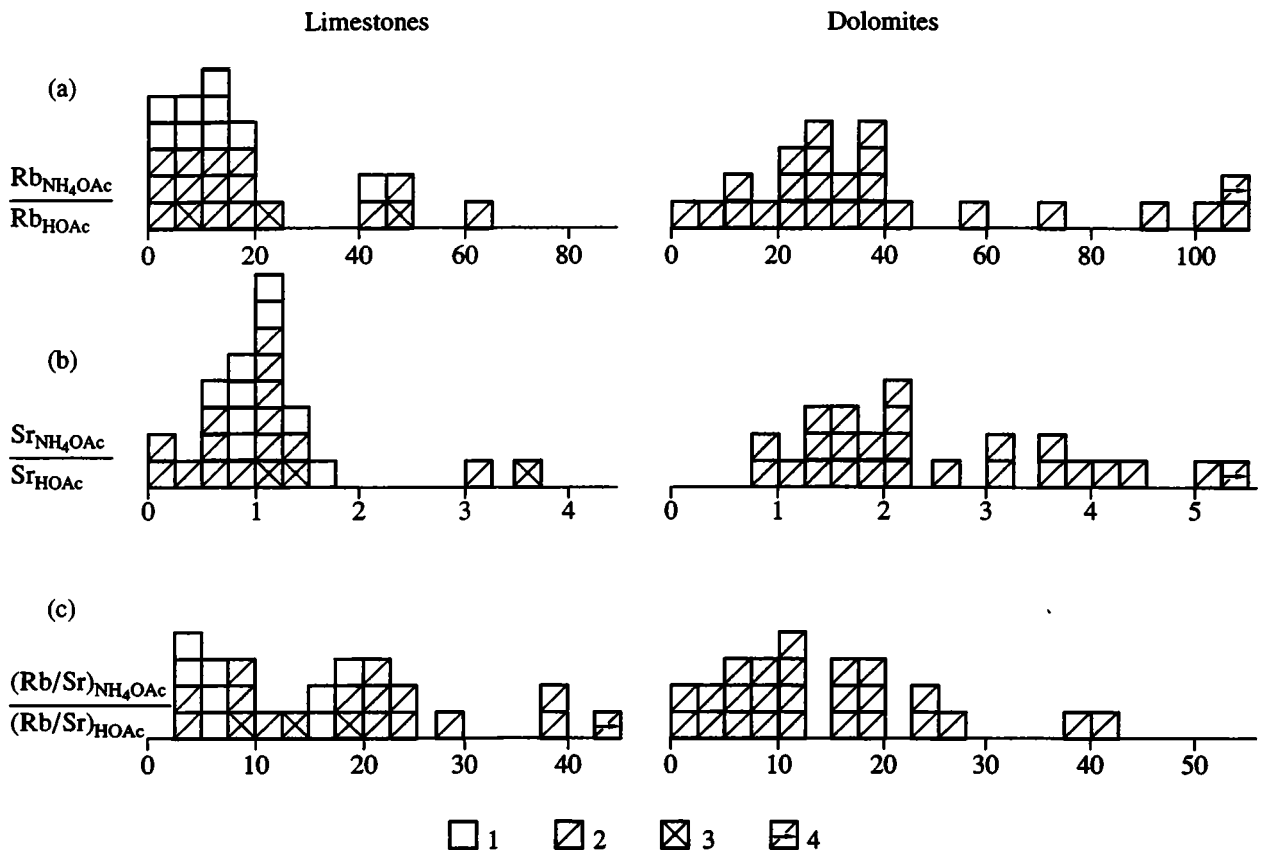

Fig. 4. Relative Rb and Sr concentations in the readily and less soluble phases of carbonate rocks marked by symbols NH_4OAc and HOAc, respectively.

1 - 3 – geographic position of samples: 1 – Ura uplift; 2 – Turukhansk uplift; 3 – Olenek uplift; and 4 – out of horizontal scale.

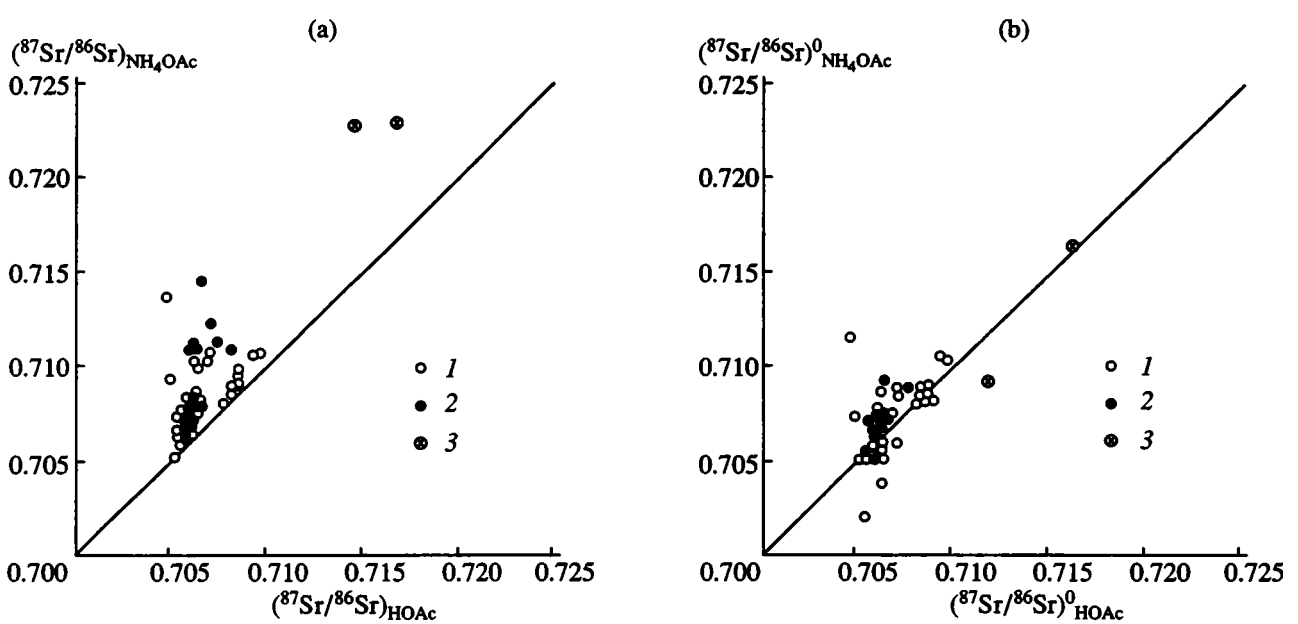

Fig. 5. Measured (a) and calculated initial (b) $^{87}Sr/^{86}Sr$ ratios in the readily (NH_4OAc) and less soluble (HOAc) phases of studied samples.

1 – limestones; *2* – dolomites; and *3* – Fe-dolomites.

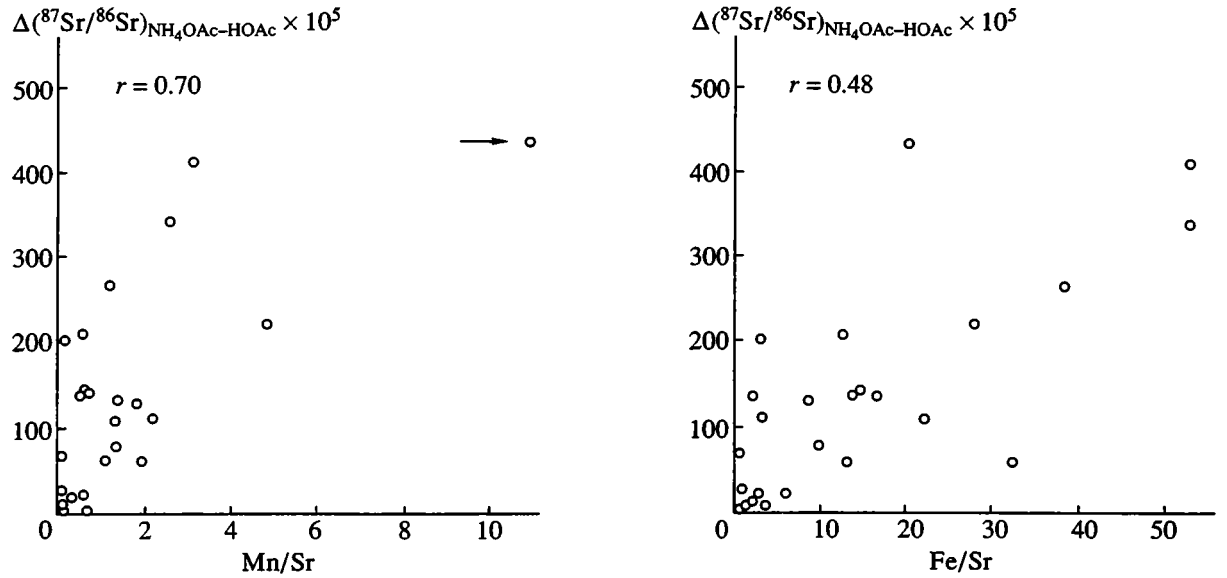

Fig. 6. Diagrams of the $^{87}Sr/^{86}Sr$ ratio difference between the readily and less soluble phases plotted versus Mn/Sr and Fe/Sr ratios in whole rock carbonate fractions of limestones (r – value of correlation coefficient).

diagenetic fluids. Comans and Middelburg (1987) have shown that the precipitation of bivalent trace elements on the calcite surface begins with monolayer adsorbtion, which is then followed by the formation of the element and calcium carbonate solid solution below this layer. The X-ray absorption spectroscopy of newly formed calcite (Pingitore *et al.*, 1992) confirmed that Sr^{2+} replaces the Ca^{2+} ions in this mineral structure as a component of the solid solution without forming a separate phase, e.g., strontianite. Thus, the limestone phase readily soluble in NH_4OAc is probably the secondary calcite composing the outer layers of mineral grains or aggregates inside pores (interstices).

Considering the studied samples, we must note the distinct positive correlation between Mn/Sr ratios in the bulk carbonate fraction of the rocks and the $\Delta(^{87}Sr/^{86}Sr)_{NH_4OAc-HOAc}$ values calculated as the difference between $^{87}Sr/^{86}Sr$ ratios in readily and less soluble phases of limestones (Fig. 6). The coefficient of this linear correlation is 0.70. Another, less evident covariance (with the correlation coefficient 0.48) characterizes the Fe/Sr and $\Delta(^{87}Sr/^{86}Sr)_{NH_4OAc-HOAc}$ values (Fig. 6). Both relations show that the intensity of disturbance of the Rb-Sr system in limestones is, to some extent, connected to their enrichment in Mn and Fe, which are the indicators of late freshwater diagenesis (catagenesis).

It is difficult to evaluate the Rb enrichment in the readily soluble phase, because we do not know the distribution coefficient of Rb^+ between the low-Mg calcite and water fluid. Nevertheless, relatively high Rb concentrations in the diagenetic phase are not surprising, as the meteoric water is able to dissolve considerable volumes of noncarbonate Rb-rich rocks when draining through them.

The revealed, apparent values of the initial $^{87}Sr/^{86}Sr$ ratios, calculated using the Rb and Sr concentrations in the readily and less soluble phases and the rock ages, are usually specific for each sample (Fig. 5b) and discriminate these phases as noncogenetic. In contrast to the measured $^{87}Sr/^{86}Sr$ ratios (compare with Fig. 5a), the apparent initial ratios do not show any regularity in distribution. In comparison with the same ratios characterizing the less soluble phase, they are relatively high in one group of readily soluble phases but lower in another.

If the distribution pattern of isotopic parameters in the studied formations is the result of a three-stage process, then the most probable succession of events should correspond to the following: deposition of carbonate sediments, their early diagenesis, and a much later stage of catagenesis (late diagenesis).

The catagenic origin of the readily soluble phase was presumably stimulated by a freshwater fluid with high Rb, Fe, and Mn concentrations, as well as $^{87}Sr/^{86}Sr$ ratios exceeding the parameters of the original limestones (the less soluble phase). This conclusion is supported by (1) a positive correlation between values of $\Delta(^{87}Sr/^{86}Sr)_{NH_4OAc-HOAc}$ plotted versus Fe/Sr and Mn/Sr ratios, and (2) imbalanced initial $^{87}Sr/^{86}Sr$ ratios, which are mostly higher in the readily soluble phase as compared to the less soluble one (see Table 3 and Fig. 5b). In the suggested geochemical environments, the inverse relation between both ratios indicates that the readily soluble phase was formed much later than the primary carbonate minerals and trapped a large quantity of Rb from the interstitial fluid. In this case, the

stratigraphic age of the rock, being mistaken for the onset of the Rb-Sr system of the readily soluble phase, should not be used for the calculation of the apparent value of its initial $^{87}Sr/^{86}Sr$ ratio. Naturally, the larger the time interval separating the sedimentation and catagenetic stages and the higher the Rb concentration in the readily soluble phase, the higher chance of obtaining a negative value of $\Delta(^{87}Sr/^{86}Sr)_{NH_4OAc-HOAc}$.

Dolomites. In the analyzed dolomites, the Mn/Sr, Fe/Sr, and Ca/Sr ratios are usually much higher than in the limestones (Table 2). This could be considered a result of the crystal chemistry, because in regard to radiuses, the Fe^{2+} (0.65 Å) and Mn^{2+} (0.91 Å) ions match Mg^{2+} (0.80 Å) better than Ca^{2+} (1.08 Å). Simultaneously, the Sr^{2+} (1.21 Å) ion is significantly larger than the Ca^{2+} and, especially, the Mg^{2+}. Consequently, the dolomites are rich in Mn and Fe but depleted in Sr, relative to calcite (Whittaker and Muntus, 1970; Kubanek and Parekh, 1976). Among the studied dolomites, the lowest Mn/Sr and Fe/Sr ratios were distinguished in rocks of the Sukhaya Tunguska Formation, whereas samples T-1, T-27 (Linok Formation), and T-48 (Derevnya Formation) revealed very high values for these ratios. Carbonates in the T-1 and T-27 samples correspond to the mineral formulas $Ca_{1.02}Fe_{0.26}Mg_{0.69}Mn_{0.03}(CO_3)_2$ and $Ca_{1.12}Fe_{0.21}Mg_{0.64}Mn_{0.03}(CO_3)_2$, respectively, thus representing either the Fe-dolomites or the calcite-ankerite mixture.

Samples T-1 and T-27 were collected from the basal part of the Linok Formation, 6 and 9 m above the boundary separating it from the underlying siliciclastic deposits of the Bezymyannyi Formation. Evidently, the ankerite component of the carbonates originated here at the expense of Fe and Mg, migrating from the underlying horizons in response to an elision process (Kholodov, 1983; Kholodov and Shmariovich, 1992), which, in addition to other effects, was responsible for Fe^{2+} and Mg^{2+} mobilization during clay dehydration. One dehydration stage, related to the illitization of smectite, could have contributed matter to the ankerite formation in overlying sediments (Boles, 1978; Boles and Franks, 1979; Kholodov, 1983). Because the Linok Formation consists of limestones with low Mg content higher in the section (usually less than 2.3% MgO, occasionally 3.0 - 3.9%), the sampled (T-1 and T-27) rocks of its basal part are the products of limestone alteration rather than the Fe-enrichment of pre-existing dolomites. This conclusion is in good agreement with data on the oxygen isotopic composition (Table 2). In addition, the Fe-rich elision waters were probably responsible for the Fe/Sr ratio increase in other rocks of the Linok Formation (see above).

Recently, Veizer *et al.* (1992) reported that $\delta^{18}O$ values in early diagenetic dolomites of the Archean–late Paleoproterozoic age interval range between −2 and −6‰ PDB. Similar or somewhat higher values should be expected for the Riphean dolomites (Fairchild and Spiro, 1987; Zempolich *et al.*, 1988; Fairchild *et al.*, 1990). However, the data in Table 2 shows that only the rocks of the Sukhaya Tunguska Formation match this range. Other dolomites experienced postdepositional alterations of various intensities. Among them, the Burovaya Formation rocks seem to display the most notable changes.

Similar to the limestones, the dolomite fraction soluble in NH_4OAc was constantly below 10%. In contrast, the summary sample portion dissolved in both NH_4OAc and HOAc was, in this case, usually lower than in the limestones, because the acetic acid is too weak to entirely dissolve the carbonate phase of dolomites. Hydrochloric acid is more effective and was usually used in the carbonate rock treatment for studying their Sr isotopic composition. However, such a treatment is also able to dissolve noncarbonate rock constituents with their own Rb-Sr systems, and many researchers (e.g., Derry *et al.*, 1989; Asmerom *et al.*, 1991) consider HCl to be unacceptable as a reagent. As to incomplete dolomite dissolving in acetic acid, it seems to insignificantly affect the results of Rb-Sr analysis of the less soluble dolomite phase. At any rate, parallel runs for limestones (Table 3) suggest that their less soluble phase is quite homogeneous in composition.

As in the limestones, the readily and less soluble dolomite phases (nonresistant against NH_4OAc and HOAc, respectively) are very different in their Rb-Sr systematics. In both rock types, Rb concentrations and Rb/Sr ratios are significantly higher in the first phase than in the second (Figs. 4a and 4c). The measured $^{87}Sr/^{86}Sr$ ratios show the same distribution trend (Fig. 5a). However, the Sr concentrations in these phases of the dolomites display an opposite situation, as compared to limestones: the readily soluble dolomite fractions usually contain more Sr than the less soluble portions (Fig. 4b). Two factors, acting separately or in combination, could be responsible for this: the different crystal chemistries (otherwise, mineralogies) of the readily and less soluble phases and the diverse chemical compositions and concentrations of fluids, which influenced the origin of both phases. The disagreement between the apparent values of the initial $^{87}Sr/^{86}Sr$ ratios, calculated using Rb and Sr concentrations in the phases, and the rock ages (Fig. 5b) suggest that the above phases are noncogenetic. It seems that most of the dolomite samples under investigation were somehow affected by the process of dedolomitization, which is related to changes in the interstitial water composition and the decrease of their Mg^{2+}/Ca^{2+} ratio (Frank, 1981; Budai *et al.*, 1984; Kashik and Mazilov, 1992).

Fresh water activity in the late diagenetic alterations of the dolomites is evident, as in the limestones, from the revealed correlation between the $\Delta(^{87}Sr/^{86}Sr)_{NH_4OAc-HOAc}$ values and the Mn/Sr or Fe/Sr ratios (Fig. 7). The correlation coefficients are 0.53 for the first case and 0.81 for the second (in both cases, data on samples T-1 and T-27 is disregarded).

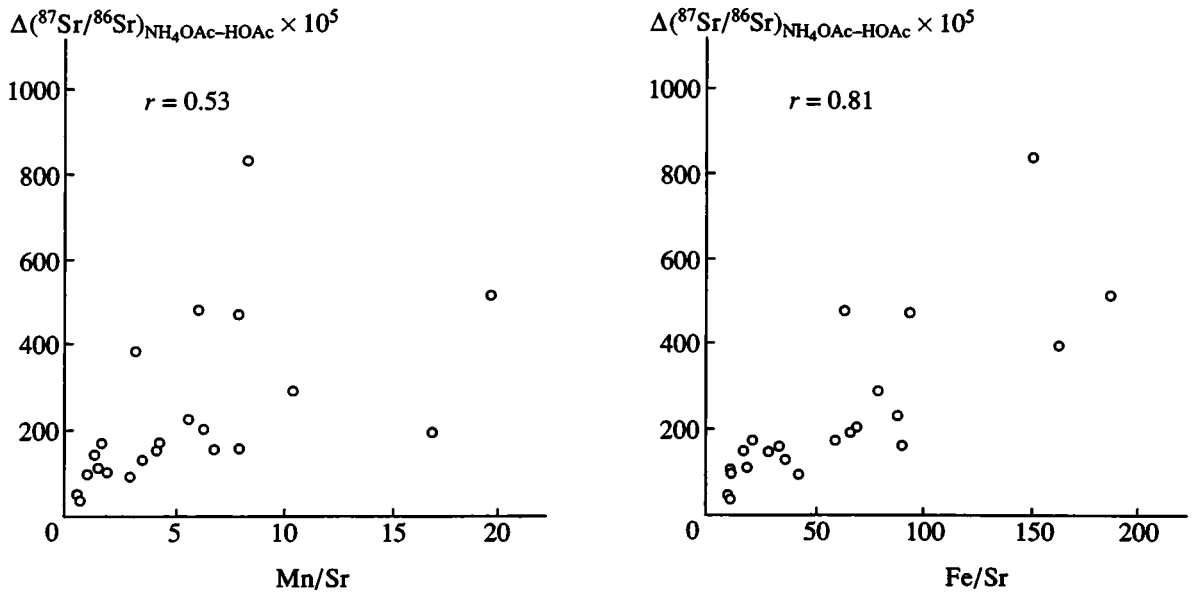

Fig. 7. Diagrams of the $^{87}Sr/^{86}Sr$ ratio difference between the readily and less soluble phases plotted versus Mn/Sr and Fe/Sr ratios in whole rock carbonate fractions of dolomites (r – value of correlation coefficient).

The $^{87}Sr/^{86}Sr$ ratios in the whole rock carbonate samples as possible indicators of the Sr *isotopic composition in the Proterozoic seawater.* As follows from the presented data, nearly all the studied rocks contain a phase (or phases) that is different in its Rb-Sr characteristics from the whole rock sample and presumably originated much later, relative to the depositional stage. This essentially limits the applicability of the whole rock carbonate samples in studies of the Sr isotopic composition in sedimentation media. Regarding dolomites, most investigators (e.g., Asmerom *et al.*, 1991; Derry *et al.*, 1992) did not consider them very promising in this respect because of their doubtful origin (e.g., Tucker, 1982, 1983; Braithwaite, 1991) and relatively low Sr concentrations. Nevertheless, most studied limestones (Brand and Veizer, 1981; Asmerom *et al.*, 1991; Derry *et al.*, 1992) were regarded, on the basis of typical geochemical criteria, as rocks lacking significant postdepositional alterations.

Recent detailed studies of Phanerozoic carbonate rocks (Moldovanyi *et al.*, 1990; Vahrenkamp *et al.*, 1991; Kupecz and Land, 1991; Montanez and Read, 1992; Barnaby and Read, 1992) revealed the coexistence of several noncogenetic generations of carbonate minerals with discordant chemical and isotopic characteristics, including their $^{87}Sr/^{86}Sr$ ratios. Undoubtedly, a similar situation is typical of many (if not all) Precambrian carbonate rocks.

The results of this study have a general methodical significance. They show that published $^{87}Sr/^{86}Sr$ ratios, which were measured in whole rock carbonate samples, potentially bear a systematic error of value depending on the range of difference between Sr concentrations and $^{87}Sr/^{86}Sr$ ratios in the noncogenetic phases. When the phases significantly differ in Sr isotopic composition, the deviation of the $^{87}Sr/^{86}Sr$ ratio in the whole rock sample from that in the primary carbonates could be registered in the fourth significant digit. This value is notable for up-to-date precision levels.

The method of preliminary leaching of carbonate rocks with NH_4OAc solution provides insight into the nature of the postdepositional alterations as well as a way to separate noncogenetic carbonate phases. The significant Rb-enrichment of the secondary (readily soluble) phase (Fig. 4a) enables us to statistically evaluate the separation degree of the phases. The very weak correlation between Rb concentrations in the readily and less soluble phases (correlation coefficient 0.28) shows that the dual procedure of leaching and ultrasonic machining results in a nearly total separation of primary and secondary carbonates. It is clear that this method significantly decreases the possibility of distorting reliable values of the initial $^{87}Sr/^{86}Sr$ ratios in limestones. For dolomites, the relevant correlation coefficient is 0.54. This means that the double treatment of dolomites with NH_4OAc solution is insufficient to entirely separate the noncogenetic carbonate phases, though it reveals the difference in their Sr isotope systems. Therefore, the $^{87}Sr/^{86}Sr$ ratios in the less soluble dolomite fraction are hardly applicable for evaluating the Sr isotopic composition of seawater.

INITIAL $^{87}Sr/^{86}Sr$ RATIOS IN THE RIPHEAN, VENDIAN, AND LOWER CAMBRIAN CARBONATES FROM SIBERIA

The $^{87}Sr/^{86}Sr$ variations in the less soluble phases of studied carbonate rocks from Siberia are shown in

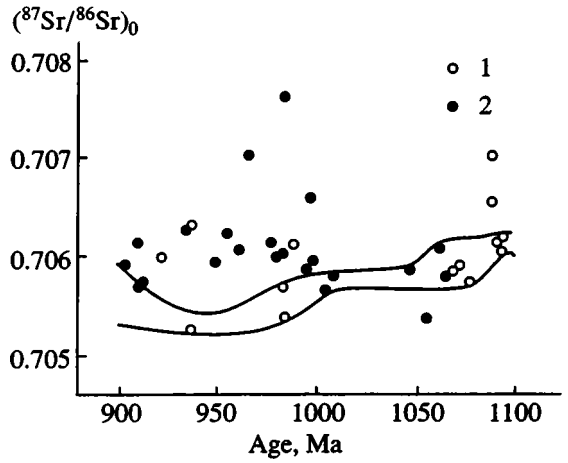

Fig. 8. $^{87}Sr/^{86}Sr$ ratio variations in the Riphean carbonate rocks of the Turukhansk uplift (lines restrict presumably unaltered samples).
1 – limestones; 2 – dolomites.

Figs. 8 and 9. Figure 9 also illustrates recently published data on the Upper Riphean and Vendian carbonates from several other regions (Veizer *et al.*, 1983; Derry *et al.*, 1989; Asmerom *et al.*, 1991; Kaufman *et al.*, 1993).

Olenek uplift. Only three samples characterizing various horizons of the succession in this region were analyzed. Therefore, the interpretation of the results is problematic. All samples have high Mn/Sr and Fe/Sr ratios (Table 2). Their readily soluble phases are rich in Rb and have notably higher $^{87}Sr/^{86}Sr$ ratios than corresponding less soluble fractions (Table 3). This seems to be the result of a process that initiated fluid migration through the siliciclastic sediments underlying the Arymas Formation, which are enriched in Mn, Fe, and Rb. In contrast, the less soluble limestone phases are Rb-depleted, and it would be admissible to consider their initial $^{87}Sr/^{86}Sr$ ratios as the minimal values characterizing seawater of the corresponding age.

Turukhansk uplift. For this region, we studied 39 samples (14 limestones, 23 dolomites, and 2 Fe-dolomites) representing the main part of the exposed Riphean succession with a time range of ~1100 to ~850 Ma.

As was noted, the very high initial $^{87}Sr/^{86}Sr$ ratios in samples T-1 and T-27 from the Linok Formation resulted from the ankerite formation after the original calcite was stimulated by fluids derived from the underlying siliciclastic rocks of the Bezymyannyi Formation. Consequently, we exclude them from Figs. 8 and 9. Nevertheless, variations of initial $^{87}Sr/^{86}Sr$ ratios in five other less soluble phases of limestone samples from this formation are rather high (0.70601 - 0.70697). Sedimentation in a basin that was influenced by a considerable continental runoff would be a plausible explanation for the generally elevated values of the initial $^{87}Sr/^{86}Sr$ ratios in the Linok sediments. Such an idea is in accordance with the suggested freshwater character of the older Bezymyannyi basin (Bazhenova *et al.*, 1970). However, the revealed difference in the $^{87}Sr/^{86}Sr$ ratios calculated for the Linok rocks is too high to be reliable for the very short interval of geological time that constrains the deposition of the unit (approximately 3 Ma, see Table 1). It is more conceivable to regard these diverse $^{87}Sr/^{86}Sr$ ratios as being related to those events that were responsible for the radiogenic ^{87}Sr-enrichment and ankerite formation in samples T-1 and T-27, near the base of the Linok Formation. This is favored by the unusually high Fe/Sr ratios in the limestones (Table 2) and the notable relative difference in $^{87}Sr/^{86}Sr$ ratios measured in their readily and less soluble phases (Table 3). Therefore, only the lowermost initial $^{87}Sr/^{86}Sr$ ratios in the less soluble phases of samples T-8, T-30, and T-31 should be considered as most representative among others from this formation.

Rocks of the Sukhaya Tunguska Formation show very limited variations in their initial $^{87}Sr/^{86}Sr$ ratios. Both the limestones and the dolomites of the formation have low Mn/Sr and Fe/Sr ratios and display $\delta^{18}O$ values that are typical for rocks of corresponding age (Table 2). The Rb concentrations in their phases of different solubility are low, and, therefore, the range of difference between measured $^{87}Sr/^{86}Sr$ ratios in these phases is small. The only exception is the dolomite T-42, in which both phases are enriched in Rb (Table 3). The abnormally low value of the "initial" $^{87}Sr/^{86}Sr$ ratio, calculated for its readily soluble phase, indicates a recent event of Rb-enrichment. The less soluble fraction of this sample is also the richest in Rb, relative to other rocks of the Sukhaya Tunguska Formation. Evidently, in this case, treatment with NH_4OAc did not result in a total separation of primary and secondary carbonates. If so, then the standard correction for the radiogenic ^{87}Sr, which accumulated in the less soluble phase during the interval corresponding to the stratigraphic age of the sample, is misleading. In other words, data on this sample should be disregarded.

Rocks of the Derevnya Formation, occurring at the base of the Upper Riphean succession of the Turukhansk uplift, have high Mn/Sr and Fe/Sr ratios (Table 2). This is especially characteristic of dolomites T-18 and T-48 with high Rb concentrations in the readily soluble phases (Table 3). Thus, it is understandable why the initial $^{87}Sr/^{86}Sr$ ratios in the less soluble phases of these samples are higher than in others. Accordingly, this indicates significant secondary alterations of the rocks.

Geochemical characteristics of rocks from the Burovaya Formation are very heterogeneous (Tables 2 and 3). In dolomites, highly variable values of Mn/Sr, Fe/Sr, Ca/Sr, and $\delta^{18}O$ are uncorrelated with either the initial $^{87}Sr/^{86}Sr$ ratios or other parameters of the Rb-Sr system. In the less soluble phase of T-57, which has an extremely low $\delta^{18}O$ value, the initial $^{87}Sr/^{86}Sr$ ratio is much higher than in other dolomites, probably because of some secondary alterations. In addition to this, the

dolomites show a rather limited range of corresponding ratios (0.70592 - 0.70627).

Limestones from the Burovaya Formation have low Mn/Sr and Fe/Sr ratios, but relatively high $\delta^{18}O$ values. This combination usually indicates an insignificant degree of postdepositional alterations. However, the initial $^{87}Sr/^{86}Sr$ ratios calculated for their less soluble phases are too variable (0.70522 - 0.70630) and chaotic to be considered as characteristic of the sedimentation medium. The least disturbed Rb-Sr system could be considered to be the one in limestone T-64. It has low Rb and high Sr concentrations in both phases as well as the lowest Mn/Sr and Fe/Sr ratios, in comparison with all studied samples. Accordingly, its initial $^{87}Sr/^{86}Sr$ ratio should be regarded as the characteristic that best approximates the seawater of that time (an estimated age of 935 Ma). It is difficult to estimate either the degree or the cause of the Rb-Sr system disturbances in other samples from the Burovaya Formation. It can only be suggested that the process of dolomitization, which somehow affected nearly the entire volume of the formation, was separated by a significant time interval from the sedimentation stage.

Dolomites from the Shorikha Formation have high Mn/Sr and Fe/Sr ratios, while their $\delta^{18}O$ values range from −6.2 to −7.4‰ (Table 2). In addition, they revealed a distinct reverse correlation between Sr concentrations and calculated initial $^{87}Sr/^{86}Sr$ ratios in the less soluble fraction (Table 3). Evidently, after formation, the dolomites were recrystallized in a medium with a high $^{87}Sr/^{86}Sr$ ratio, and the process affected both the readily and the less soluble phases. The minimal $^{87}Sr/^{86}Sr$ ratio in the fluid that initiated the recrystallization can be estimated using the highest modern value of the ratio in the readily soluble dolomite phase (0.70798, Sample T-119). In the sedimentation (dolomitization?) medium, the minimal $^{87}Sr/^{86}Sr$ ratio should presumably correspond to the lowest initial value in the less soluble phase (0.70569, Sample T-114).

According to geochemical characteristics (Table 2), the single analyzed sample of the Turukhansk Formation seems less altered than dolomites of the Shorikha Formation. The initial $^{87}Sr/^{86}Sr$ ratio in this sample is somewhat lower than in carbonate rocks of similar age from northwestern Canada (Asmerom *et al.*, 1991).

Ura uplift. We studied nine samples from this region. Six of them are from the upper Upper Riphean succession (the Nikol'skaya and Chencha formations); one sample is from the Middle Vendian section (the Tinnaya Formation); and two represent a lower part of the Lower Cambrian (the Macha and Tolbachan formations). All the rocks are limestones.

The initial $^{87}Sr/^{86}Sr$ ratios in the less soluble phases of limestones from the Nikol'skaya and Chencha formations range from 0.70811 to 0.70939 (Table 3), i.e., they are considerably higher than the values (0.70662 - 0.70727) measured in the coeval carbonates from Spitzbergen and East Greenland (Derry *et al.*, 1989, 1992). As the $^{87}Sr/^{86}Sr$ ratio variations in geologically coeval oceans should be below 3×10^{-5} (Capo and DePaolo, 1992), the above fact definitely suggests that the Sr isotopic composition of rocks from the Nikol'skaya and Chencha formations is inappropriate as a characteristic of the seawater of the Late Riphean oceans.

The Mn/Sr ratios in limestones of the Nikol'skaya Formation (Table 2) are moderate, though higher on average in comparison with many other analyzed limestones. In the Chencha Formation carbonates they are low, but this seems to reflect only the high Sr concentrations in the oncolite and stromatolite limestones of this formation (Table 3), which were originally composed of aragonite (Golovenok, 1985).

The oxygen and carbon isotopic composition in analyzed samples of the Nikol'skaya and Chencha formations has not been investigated. In general, the $\delta^{18}O$ and $\delta^{13}C$ values known for these formations are anomalously low. In the Nikol'skaya Formation, they range from −7.8 to 12.4‰ PDB and from −5.4 to −12.8‰ PDB, respectively. In the Chencha Formation, the corresponding ranges are −7.1 to −10.2‰ PDB and −7.6 to −8.4‰ PDB (Pokrovskii and Gertsev, 1993). The cited authors believe that such parameters indicate large-scale catagenetic alterations of the sediments. Large celestite crystals and secondary cement of the Chencha oncolite limestones (Golovenok, 1985) also suggest intense fluid circulation through the Upper Riphean carbonate rocks of the Ura uplift.

The single analyzed sample from the Tinnaya Formation of Middle Vendian age revealed a somewhat high Fe/Sr ratio and a very low $\delta^{18}O$ value (Table 2). According to Pokrovskii and Gertsev (1993), the carbonates of this formation also suffered catagenetic recrystallization, though to a lesser extent than underlying rocks of the Zhuya Group. Undoubtedly, this process also affected Sample 980/100. Respectively, its less soluble fraction probably differs from the Vendian seawater in its Sr isotopic composition.

The Lower Cambrian limestones of the Macha and Tolbachan formations showed high Rb concentrations in the readily soluble phases (Table 3), but other geochemical criteria (Table 2) do not favor any kind of evident secondary alterations. Therefore, we may suppose that the Sr isotopic composition in less soluble phases of samples 980/140 and 980/151 could be characteristic of the primary carbonate minerals.

Thus, the results of the integrated geochemical and isotopic studies on the Riphean, Vendian, and Cambrian carbonates from Siberia allow us to select, though ambiguously, those analyzed samples that best approximate the Sr isotopic composition of the sedimentation medium. For the interval 0.5 - 1.1 Ga, we outlined the corresponding pathway of initial $^{87}Sr/^{86}Sr$ ratios in Figs. 8 and 9, using recently published data on other regions in addition to our results (Veizer *et al.*, 1983; Derry *et al.*, 1989, 1992; Asmerom *et al.*, 1991; Kaufman *et al.*, 1993). To confine the pathway variations, we

always considered samples that were discriminated on the basis of the above geochemical criteria. The samples that do not meet these criteria usually have higher values of the initial $^{87}Sr/^{86}Sr$ ratios in the less soluble phase.

Our results for the Lower and lower Middle Riphean of the Olenek uplift are too fragmentary to provide far-reaching conclusions. Still, they suggest that in the seawater of that age, the Sr isotopic composition was considerably less radiogenic than has usually been believed (Veizer and Compston, 1976; Veizer *et al.*, 1983; Faure, 1986), and it showed a clear trend of radiogenic Sr increasing with time. Pokrovskii and Vinogradov (1991) also reported low initial $^{87}Sr/^{86}Sr$ ratios in the Lower and Middle Riphean carbonates from the Anabar massif. However, their data, which was obtained on dolomites without a preliminary evaluation of secondary alterations, are difficult to interpret.

The results for the younger Riphean horizons of the Turukhansk uplift are more representative, as compared to those for the Olenek region. In general, they agree with the known trend of the Sr isotopic variations in seawater of that time but displace the variation curve down to a field of lower values. New data indicates that 1080 - 1100 Ma ago the $^{87}Sr/^{86}Sr$ ratio in seawater was below 0.70601 - 0.70620 (samples T-8 and T-30 from the Linok Formation). The analyzed rocks from the Sukhaya Tunguska Formation, which crowns the Middle Riphean succession of the Turukhansk uplift, show a tendency for further ratio decrease. This tendency is spotted by data on the least altered sample (T-52) from the Derevnya Formation, which is the basal one in the Upper Riphean succession of the region. Subsequently, the least altered limestones from the overlying Burovaya Formation show a notable decrease in the $^{87}Sr/^{86}Sr$ ratios to the range 0.70522 - 0.70537. The estimated age of these limestones is within the interval 935 - 984 Ma (Table 1). Therefore, this depression on the $^{87}Sr/^{86}Sr$ variation curve closely matches a similar depression distinguished by studying North African materials ("The Mantle Event" at ca. 900 Ma ago, Veizer *et al.*, 1983) and has a somewhat greater amplitude (0.70522 - 0.70537 against 0.7056 - 0.7074). As the younger (850 - 900 Ma) deposits in both the Turukhansk region (this work) and northwestern Canada (Asmerom *et al.*, 1991) yielded higher values of the initial $^{87}Sr/^{86}Sr$ ratios, their decrease within the interval 920 - 980 Ma (Figs. 8 and 9) seems reliable.

The initial $^{87}Sr/^{86}Sr$ ratios in the least altered dolomites from the upper horizons of the Turukhansk succession (the Shorikha and Turukhansk formations) agree well with their values in the coeval and younger deposits of other regions (Derry *et al.*, 1989, 1992; Asmerom *et al.*, 1991). This information indicates a quick increase of the $^{87}Sr/^{86}Sr$ ratios during the second half of the Late Riphean.

Studied samples of the upper Upper Riphean succession, i.e., those from the Nikol'skaya and Chencha formations of the Ura uplift, contain highly radiogenic Sr (probably because of secondary alterations) and, thus, drop out of the general trend of $^{87}Sr/^{86}Sr$ ratio variations during this period (Fig. 7). As to the Middle Vendian sample from the Ura uplift, its $^{87}Sr/^{86}Sr$ ratio in the less soluble phase is very similar to the ratio values in the Vendian samples from other regions (Kaufman *et al.*, 1993). This may indicate that, despite some geochemical anomalies (see above), the postdepositional alterations that affected this limestone were not very intense. Two samples from the lower Lower Cambrian Macha and Tolbachan formations of the Ura uplift revealed somewhat lower $^{87}Sr/^{86}Sr$ ratios in comparison with the South African samples of similar age (Kaufman *et al.*, 1993). With these limited results, it is difficult to evaluate their general significance, as the decrease may indicate either less radiogenic (as compared to present opinions) Sr isotopic composition in the Early Cambrian seawater or some particular fluctuations against the background of a main ascending trend.

CHEMOSTRATIGRAPHIC AND PALEOGEODYNAMIC SIGNIFICANCE OF THE RESULTS

Recent studies (Derry *et al.*, 1989, 1992; Asmerom *et al.*, 1991; Kaufman *et al.*, 1993) revealed data of up-to-date precision that characterizes the Sr isotopic composition in carbonates of the latest Proterozoic age, i.e., in those of the Vendian and late Late Riphean. This work presents similar results for an older rock package comprising the lower Upper and upper Middle Riphean horizons. The pertinent succession of the Turukhansk uplift was used as a model for obtaining the necessary data. As to the results concerning the upper part of the Siberian Upper Riphean, they have been obtained from distinctly altered samples and are unsuitable for consideration, while the data on rocks of the Vendian–Lower Cambrian (Ura uplift) and Lower-lower Middle Riphean (Olenek uplift) are very fragmentary.

The results on the Turukhansk succession match well with the data on younger carbonates (Derry *et al.*, 1989, 1992; Asmerom *et al.*, 1991; Kaufman *et al.*, 1993), and in Fig. 9, they directly continue the pertinent variation curve of the Sr isotopic composition in seawater down to the field of more ancient deposits. In addition, this confirms that upper horizons of the Upper Riphean (younger than 850 - 800 Ma) do not occur in the Turukhansk region. Such a conclusion was originally suggested on the basis of interregional correlations and scarce data of isotopic geochronology (for interpretation and references, see Semikhatov and Serebryakov, 1983) and is now supported by data on carbon isotope chemostratigraphy (Knoll *et al.*, in press).

From the mid-Late Riphean to the terminal Vendian, the general increase of initial $^{87}Sr/^{86}Sr$ ratios in carbonates is comparable to that in seawater of the period that has lasted from the mid-Mesozoic until the Recent.

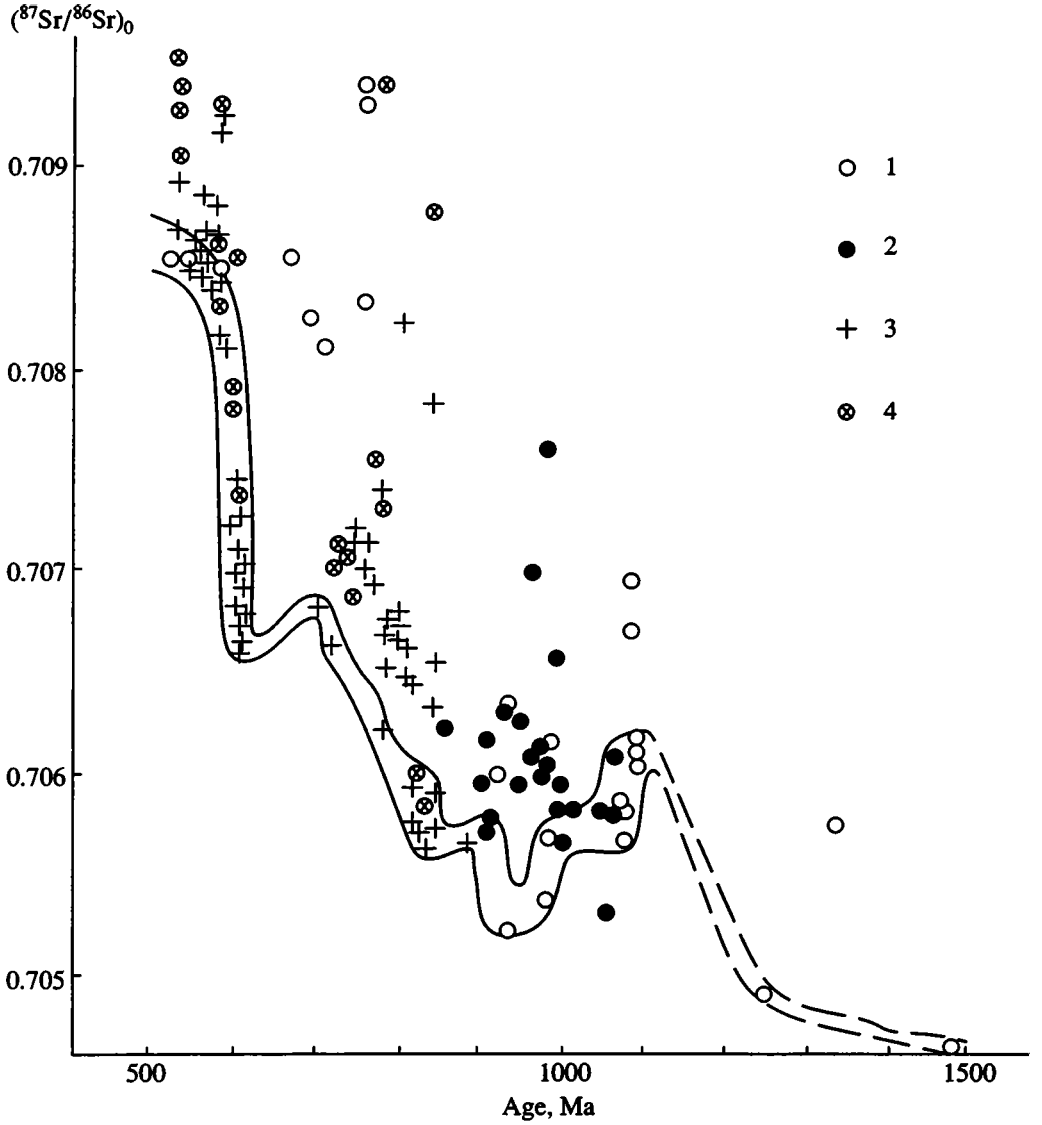

Fig. 9. $^{87}Sr/^{86}Sr$ ratio variations in the Late Proterozoic and Early Cambrian seawater (lines restrict most reliable values characterizing relevant intervals).

1, 2 – data from this work: 1 – limestones; 2 – dolomites; 3, 4 – data from published works (Veizer *et al.*, 1983; Derry *et al.*, 1989; Asmerom *et al.*, 1991; Kaufman *et al.*, 1993): 3 – limestones; 4 – dolomites.

According to data available for the interval 930 - 580 Ma, the estimated mean rate of the ratio growth is 1.3×10^{-4} per 10 Ma. As this value is much higher than the analytical precision of modern mass spectrometers, the method of Sr isotopic stratigraphy (DePaolo and Ingram, 1985; Faure, 1986) with an unbelievable (for the Precambrian) resolution of ±5 Ma could be applicable for the Late Riphean–Early Vendian time span if the $^{87}Sr/^{86}Sr$ ratio variations were unidirectional and monotonous. However, there is no doubt that for the Late Riphean–Vendian interval the variation curve of the $^{87}Sr/^{86}Sr$ ratio in seawater has different slopes and inflection points (e.g., Derry and Jacobsen, 1988; Asmerom *et al.*, 1991). Therefore, the high resolution potential of Sr isotopic stratigraphy of the Late Precambrian is of limited applicability at present, because the standard curve of Sr isotopic variations in the Proterozoic seawater needs to be known in more detail and better calibrated against the numerical time scale. Such a curve must be based on congruent data characterizing several geographically widely separated and well-dated sections of the same age. Nevertheless, even now the available information on the initial $^{87}Sr/^{86}Sr$ ratio variations in the Proterozoic carbonates allows Sr isotopic stratigraphy to be useful for the correlation of the Upper Riphean–Vendian deposits within the known chronostratigraphic and chronometric scales of the Precambrian.

Since the Sr isotopic composition of seawater depends, like many other geochemical characteristics,

on global but not local factors, secular variations of $^{87}Sr/^{86}Sr$ ratios in oceans are interesting from a paleogeodynamic viewpoint. In the geological past, this ratio ranged between values typical of the continental (Fc) and mantle (Fm) fluxes. Continental river runoff is probably a nonlinear function of the total continental area size, while the mantle flux depends on the intensity of ocean water circulation through sea-floor basalts and is considered to be proportional to mantle heat generation and dissipation (Veizer, 1988; Richter *et al.*, 1992). Therefore, in the early history of the Earth, when continents were small in contrast to the high rate of heat generation in the mantle, the mantle flux should be the main factor of $^{87}Sr/^{86}Sr$ fluctuations in seawater. Later, along with continental growth and decreasing heat generation in the mantle, the continental flux gradually became dominant. The Proterozoic seems to be a transitional period between these extreme situations (Derry and Jacobsen, 1988).

The main secular trend of increasing $^{87}Sr/^{86}Sr$ ratios in seawater results from continental area increase, provenance aging, and declining mantle input into the general Sr budget of the oceanic reservoir. Episodic variations superimposed on this trend reflect less significant factors, such as the size and relief of continental provenances, lithologic differences of crustal rocks under weathering, climatic changes, and variations of volcanic activity. Concrete events and processes are responsible for the growth or decline of $^{87}Sr/^{86}Sr$ ratios during short intervals commensurate to the Sr residence time in oceanic water (~2.5 Ma, Hodell *et al.*, 1990) or to somewhat longer episodes.

The rate of radiogenic ^{87}Sr input into oceans may increase because of (1) sea level lowering, which exposes and subjects to continental weathering new land areas with ancient crustal rocks and the products of their hypergene alterations; (2) orogeny, which involves larger volumes of crustal rocks in denudation, especially the ancient formations that have risen to high hypsometric levels; and (3) more intense weathering stages, which are related either to changes in CO_2 and O_2 concentrations in the atmosphere or global climatic variations (glacial or humid epochs). The input increase may reflect both the greater continental runoff and the higher $^{87}Sr/^{86}Sr$ ratio in this flux. Richter *et al.* (1992) have shown that these two factors had nearly similar effects in respect to the $^{87}Sr/^{86}Sr$ ratio increase in the post-Eocene ocean.

As to the influence of sea regression on the Sr isotopic composition in oceans, the present opinions are controversial. DePaolo (1986) suggests that regressions caused the $^{87}Sr/^{86}Sr$ ratio to increase in seawater because of the areal growth of continental surface being affected by weathering and because of greater carbonate compensation depth. The latter decreases the volume of dissolving sea carbonates and, consequently, the diagenetic reflux of nonradiogenic Sr (Fs) into oceans. Ludwig *et al.* (1988) have actually shown that large Cenozoic regressions and related stratigraphic hiatuses were associated with enlarged continental Sr flux and higher $^{87}Sr/^{86}Sr$ ratios in seawater. However, Palmer and Edmond (1992) consider weathering reactions in floodplains to be an ineffective source of Sr, which can be extracted from river waters by soil salts precipitating in these plains. These researchers believe that by exposing mainly continental shelves the regressions are unable to result in a considerably greater Sr input through global river discharge, and that they may even sometimes cause its decrease.

Raymo *et al.* (1988) suggested that the Himalayas and Tibet, which were highly uplifted and, thus, subjected to a quick erosion, are responsible for the $^{87}Sr/^{86}Sr$ ratio increase in seawater during the last 5 Ma. In fact, the drainage river system of the region bears a considerable amount of the total Sr migrating into oceans (Palmer and Edmond, 1989; Richter *et al.*, 1992). Richter and colleagues also presented evidence that the initiation of greater Sr flux with higher $^{87}Sr/^{86}Sr$ ratios in continental runoff was coeval to the collision between India and Eurasia. According to their evaluation, the amount of Sr mobilized after the collision was large enough to account for the $^{87}Sr/^{86}Sr$ ratio evolution in Cenozoic oceans. It is also important that the mean global rate of carbonate sedimentation during the last 40 Ma has been growing parallel to the additional flux of Sr (and of Ca, accordingly) via the Himalayan rivers (Southam and Hay, 1981). In addition, the period 15 - 20 Ma ago, when the Himalayas and Tibet experienced very quick erosion, strictly corresponds to the time of a sharp change in the seawater Sr isotopic composition (Richter *et al.*, 1992).

As suggested by many investigators (Armstrong, 1971; Berner *et al.*, 1983; Berner and Rye, 1992; Sochava, 1992), the intensification of continental weathering and the related growth of the $^{87}Sr/^{86}Sr$ ratio in seawater could be related to changes in the Earth's surface temperature and atmosphere composition, being independent of continental relief. In particular, ice sheets of the temperate latitudes could potentially be responsible for dissolving the rocks below them by means of thaw water action (Armstrong, 1971; Hodell *et al.*, 1990). A large volume of sediments, continuous water aeration by turbulization, and the mechanical disintegration of denuded crystalline rocks accelerate chemical weathering. The glacio-eustatic lowering of the sea level and the related expansion of erosion areas are indirect consequences of glaciations influencing the continental Sr flux (see above), but the relative role of both factors in adding the dissolved Sr to the total flux is insufficiently clear and probably insignificant (DePaolo, 1986; Kaufman *et al.*, 1993). Similarly, it is unclear to what extent climate humidization and related intensification of weathering are able to affect the migration of radiogenic ^{87}Sr into oceans, though the mechanism itself seems plausible (DePaolo, 1986).

The decelerated input of radiogenic ^{87}Sr into oceans could be related to (1) epochs of global transgressions reducing land areas subjected to weathering; (2) acceleration of sea-floor spreading, which probably resulted in higher hydrothermal activity; and (3) periods of intense accretion of younger crustal formations along active continental margins, which changed the composition of provenances.

The decelerated rate of $^{87}Sr/^{86}Sr$ ratio growth in seawater during the transgression epochs was detected by the analysis of Neogene carbonates (DePaolo, 1986). Here, it is important that we deal with the rate deceleraton but not with the absolute ratio decrease. Thus, despite the ability to reduce the continental input of radiogenic ^{87}Sr into seawater, sea level increase during the Cenozoic apparently was not the factor that was able to reverse the tendency of a sharp growth of $^{87}Sr/^{86}Sr$ ratios in oceans. Therefore, it would be senseless to search for the correlation between the sea level position and the absolute values of the $^{87}Sr/^{86}Sr$ ratios in oceanic water (see Veizer, 1988).

Oceanic crust origin in sea-floor spreading zones is accompanied by the intense, heat-induced circulation of seawater through newly formed basalts down to a depth of 1 - 2 km (Spooner *et al.*, 1977; Hart and Staudigel, 1982). The Sr concentration in seawater remains therewith nearly constant, and the hydrothermal circulation affects only the Sr isotopic composition. The duration of this chemically active circulation phase, which is responsible for the Sr isotope exchange between basalts and seawater, is less than 5 - 10 Ma (Hart and Staudigel, 1982), and the mantle flux of nonradiogenic Sr (Fm) appears to be proportional to the spreading rate (Holland, 1984).

The $^{87}Sr/^{86}Sr$ ratio in river waters of recent island arcs and active continental margins varies from 0.7045 to 0.7085 (Goldstein and Jacobsen, 1987). In rivers of the Philippines, its range is 0.7045 - 0.7063, which corresponds to the Sr isotopic composition of recent volcanic rocks of the region. In Japanese and east Australian rivers, the $^{87}Sr/^{86}Sr$ ratios are somewhat higher (0.7063 - 0.7084), indicating the presence of older continental formations inside the relevant drainage systems.

Turning now to the variation curve in Fig. 9, we must note the gradual growth of $^{87}Sr/^{86}Sr$ ratios during the Early and early Middle Riphean. This growth is in accordance with geological data on the generally high freeboard of the Earth's continental block of this period, especially during the Early Riphean (Semikhatov, 1974; Khain and Bozhko, 1988). These variations are hardly monotonous, yet we need to discuss them in more detail with a much larger amount of experimental data.

The seawater $^{87}Sr/^{86}Sr$ ratio growth up to the range of 0.7057 - 0.7062, which is related to the period 1100 - 1070 Ma, apparently resulted from the intensification of continental erosion induced by the Grenville orogeny and the formation of the Rodinia supercontinent (Powell *et al.*, 1993). The interval of geological history 990 - 930 Ma ago was marked by an extremely high hydrothermal flux, resulting in a decrease of the $^{87}Sr/^{86}Sr$ ratio to 0.7052 - 0.7054. This interval would probably be longer if the value of 0.70562 for carbonates of the West African Adrar Formation, estimated to be approximately 890 Ma old (Veizer *et al.*, 1983; Derry *et al.*, 1989), was also considered. The significant hydrothermal flux could result from configuration changes of divergent and transform boundaries between oceanic plates (Owen and Rea, 1985). In addition, the Late Riphean was a time of widespread sea transgression (Keller *et al.*, 1974; Khain and Bozhko, 1988), which was probably related to the volume growth of mid-ocean ridges and partially responsible for the decreased continental flux of radiogenic ^{87}Sr into oceans.

The next increase in $^{87}Sr/^{86}Sr$ ratios during the second half of the Late Riphean could be related to the influence of the Jinning or early Baikalian folding 850 - 800 Ma ago and is probably related to extensive glaciation during the African glacioera 850 - 740 Ma ago (Chumakov, 1978). The considerable glacio-eustatic lowering of the sea level and the related draining of vast land areas composed of older sediments and ancient crustal complexes would be additional factors that mobilized the radiogenic ^{87}Sr flux via rivers.

The breakup of the Rodinia supercontinent was initiated 700 Ma ago (Powell *et al.*, 1993), and new passive continental margins totaling over 20 000 km in length were formed shortly after (Bond *et al.*, 1989). Nevertheless, the $^{87}Sr/^{86}Sr$ ratio in seawater of that time was somewhat lower than before, which could be related to oceanic crust generation in rift basins. Tectonic processes of the time under consideration forced apart blocks of continental crust, giving rise to new oceans: the Central Asian, Iapetus, and others (Bond *et al.*, 1984; Khain and Bozhko, 1988). In addition, they induced subsynchronous processes of subduction, e.g., in the Avalon–Cadomian belt (Murphy and Nance, 1991), and abundant collision events, which formed the vast Panafrican orogenic belt and Gondwanaland, the new supercontinent comprising the fragments of the previous one except the Baltica, Siberia, and Laurentia.

The high spreading rate, inferred for the Vendian on the basis of paleomagnetic data on the Baltica continent (Torsvik *et al.*, 1992), should cause an acceleration of nonradiogenic Sr flux into seawater at the expense of both the hydrothermal sources and the erosion of basic volcanites in active continental margins. This presumes a lowering of the $^{87}Sr/^{86}Sr$ ratio in seawater; yet, it was quickly growing instead (Fig. 9), which indicates a higher rate of continental Sr flux into the Vendian oceans. Comparable tendencies of $^{87}Sr/^{86}Sr$ ratio increase in the Vendian and Cenozoic oceans suggest that continental collisions of the Rodinia fragments resulted in the Panafrican orogeny of a scale similar to that in Himalayas. Consequently, this orogeny rose

large masses of metamorphic and intrusive rocks to the surface, which were formed from the material of ancient continental crust. Being subjected to erosion, they caused the greater continental Sr flux, which is characterized by higher $^{87}Sr/^{86}Sr$ ratios.

The considerable growth of $^{87}Sr/^{86}Sr$ ratios in seawater during the period 620 - 550 Ma indicates that proportions between the continental and mantle Sr fluxes (Fc/Fm) of that time were distinctly in favor of the former. It is all the more remarkable because the mantle flux of nonradiogenic Sr was, as noted, very intense during the Vendian. Therefore, we have to admit that the Vendian is a unique period of the Earth's history in respect to the radiogenic ^{87}Sr input into oceans. In addition to those factors related to the relief of Vendian continents, the continental Sr flux of that time could be a consequence of the intense Early Vendian glaciation (Chumakov, 1978, 1992) and the high weathering rate during the Late Vendian. The high kaolinite content in the Upper Vendian shales of the Russian platform (Ronov *et al.*, 1990) clearly favors the latter factor.

CONCLUSION

All studied limestones and dolomites of Late Proterozoic and Early Cambrian age contain a carbonate phase readily soluble in NH_4OAc, probably represented by calcite of late diagenetic origin. This phase usually reveals considerable Rb concentration and elevated $^{87}Sr/^{86}Sr$ ratios. The formation of such a phase was followed by rock enrichment in Mn and Fe.

Because of a significant difference between the primary and secondary carbonates in their $^{87}Sr/^{86}Sr$ ratios, the whole rock samples should not be used to study the Sr isotopic evolution in paleooceans. Leaching with ammonium acetate provides a better understanding of the postdepositional alterations affecting the Rb-Sr systems of carbonate rocks as well as a method for separating or considerably concentrating the noncogenetic mineral phases. This method is effective when applied to limestones and less useful in studying dolomites.

The variation curve of the $^{87}Sr/^{86}Sr$ ratio in seawater of the late Precambrian shows a distinct minimum within the interval 990 - 890 Ma. The obtained data better constrains this period, when mantle flux dominated over continental runoff and provided oceans with a greater input of nonradiogenic Sr. In addition, different proportions of continental and mantle fluxes into Late Riphean and Vendian oceans suggest significant changes in global geodynamics at the boundary between these two subdivisions of the Late Proterozoic.

Comparable amplitudes of $^{87}Sr/^{86}Sr$ ratio growth in Late Riphean–Vendian and Cenozoic seawater provide the possibility of using Sr isotopic chemostratigraphy in a wide interval (over 300 Ma) of the late Precambrian.

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New Data on Stratigraphy of Permian–Triassic Volcanogenic Deposits in the Central Part of the Tunguska Syneclise

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Abstract – Volcanogenic deposits of the Tunguska syneclise are rich in fossil faunal and plant remains. As a consequence, the first lithostratigraphic scheme that was suggested for the deposits in the central part of the syneclise (Sadovnikov, 1960, 1974) was used to define regional stratigraphic units of the Upper Permian in the Angarida (*Resheniya* ..., 1981). The imperfection of this scheme now impedes further progress in local stratigraphy. We revised the main sections of volcanogenic deposits along the Nizhnyaya Tunguska River, which earlier served as the basis for the original stratigraphic scheme. New data and the revision of previous results enabled us to distinguish many stratigraphic details of the central Tunguska syneclise as well as to suggest a new correlation between units of the Korvunchana Group, occurring in the axial part, and units from southwestern to eastern sectors of the syneclise limbs. The basal units of the Korvunchana Group, which were assigned somewhat doubtfully to the Tutonchana Horizon, occur only in the axial syneclise part. The overlying deposits, previously referred to the Dvurogii Horizon, are subdivided into the Lebedev and Khungtukun horizons. These are more widespread and overlie the Gagarii Ostrov Horizon in the syneclise limbs.

In 1929, S.V. Obruchev distinguished tuffaceous deposits of the Tunguska syneclise as representing a separate stratigraphic unit: the tuffite group. Later, in 1935, Shorokhov suggested an initial stratigraphic scheme that subdivided these deposits into lower "breccioid tuffites" and overlying "sandstones and shales" (Sadovnikov, 1974). Considering the coal-bearing deposits below the breccioid tuffites as the separate Uchami Formation of the Korvunchana Group, Shorokhov (1974a) thus recognized a third, lower unit of volcanogenic rocks (that between the tuffites and the Uchami Formation). In addition, Shorokhov (1974b) distinguished the Tura Formation, consisting mainly of basalts with subordinate tuffaceous and sedimentary rocks. The possibility of dividing the volcanogenic deposits of the Korvunchana Group into dominantly tuffaceous and tuffosedimentary local units is well-approved, and such units are widely used in geological mapping of various scales in the Tunguska syneclise region. However, it was realized that this succession is divisible into more than three members. On the basis of drilling results, Zav'yalov suggested it to be four-membered (Smirnov, 1968), while Belozerov recognized five members (*Geologicheskaya Karta* ..., 1985; *Resheniya* ..., 1981). During 1987 - 1988, we reinvestigated tuffaceous deposits exposed along the Nizhnyaya Tunguska River between two tributaries, the Iritka and Nizhnyaya Bugarikta rivers. Below, we present our data based on new observations and revised drilling results.

The most complete and best studied succession of the Korvunchana Group is exposed in the inner part of the southern limb of the Tunguska syneclise, where it approaches the axial zone between the Nidym and Taimura river mouths. Here, the lower unit consists of tuffosedimentary deposits (0 - 60 m) without paleontological remains. According to Zav'yalov, the deposits do not represent a separate unit, yet Boruchinkina *et al.* (1974) and Belozerov (*Geologicheskaya Karta* ..., 1981, 1985) considered them correlative to the Tutonchana Formation. The gravelly xenotuffs overlying these deposits and the grey, coal-bearing deposits of the Upper Permian were distinguished by Zav'yalov as the Allyun Formation. Mogucheva (1973) defined them as bearing *Madygenia* sp. forms. Near Nidym Settlement (Well 7p), the deposits are over 400 m thick. About 14 km westward, near Babkino Settlement (Well 1p), their thickness decreases to 175 m and then to 90 m near the Eksa Cliff (Well 22p, 66 km west-southwest from Well 7p). In the lower course of the Taimura River, upstream from the Dyuke Cliffs, similar tuffs form the base of the Korvunchana Group and rest on bituminous siltstones of the Gagarii Ostrov Horizon. Here, they are only 20 - 30 m thick (Boruchinkina *et al.*, 1974), and it is likely that southward and westward, the Allyun tuff beds pinch out until they finally disappear.

Tuffosedimentary rocks bearing abundant plant and animal remains overlie the Allyun Formation of the lower course of the Taimura River and have been recognized as the Dyuke Beds, with a stratotype locality in corresponding cliffs (Boruchinkina *et al.*, 1974). We suggest that this unit is of the formation rank. It is 10 - 100 m in thickness and underlies the gravelly, dominantly vitroclastic tuffs, which are 45 - 190 m thick. We propose to define the latter as the Chomchorokon Formation, whose stratotype is on the right

side of the Nizhnyaya Tunguska River, opposite the Chomchorokon River mouth. Here in the Eksa Cliffs, an upper part of the formation is exposed at the base of the coastal slope and consists of grey gravelly tuffs with columnar jointing (6 m). Below (after an unexposed interval of 10 m), its lower horizons are composed of grey ash tuffs with rare larger clasts and pisoliths. Well 22, in a middle part of the exposure, penetrated the formation at a depth interval of 36 - 64 m. The beds' succession, resting on the Chomchorokon Formation of the Eksa Cliffs, is the following:

1. Ash tuffite, soft, black, pisolith-bearing. Thickness: 5 m.

2. Greenish-grey, pisolith-bearing tuffite with inclusions (bombs?) of grey ash tuffite, also pisolith-bearing. Thickness: 3.5 m.

3. Aleuritic tuff of variable grain size: coarse at the base (brown, fine-grained tuffstones with lensing cross-bedding); dark grey and fine-grained in the middle part; and again coarse, greenish-grey, thin-platty at the top. Thickness: 5.5 m.

4. Silty tuffite, black, fine-grained, pisolith-bearing. Thickness: 5.5 m.

5. Aleuritic tuff, deep greenish-grey, fine-grained, enclosing interlayers of greenish-grey silty tuffite and coarse-grained aleuritic tuff. Thickness: 3.5 m.

6. Ash tuffite, black, with pisoliths. Thickness: 3.5 m.

7. Aleuritic tuff, brown and cherry-red, coarse- to fine-grained. Thickness: 6 m.

8. Silty tuffite with pisoliths, greenish-black, fine-grained, alternating with reddish-brown, fine- to coarse-grained aleuritic tuff. Thickness: 8 m.

9. Aleuritic tuff, brownish-red, fine-grained. Thickness: 1.5 - 5 m.

Belozerov (Boruchinkina *et al.*, 1974; *Geologicheskaya Karta* ..., 1981; *Resheniya* ..., 1981) considered these deposits, which contain abundant forms of *Parajacutiella angusta* Mog., as the Eksa Beds, but we suggest attributing them to formation rank. It should be noted that the Chomchorokon tuffs were included earlier in the Allyun Formation, while the Dyuke Beds were either considered to be a lens inside the Allyun tuffs or were mistakenly correlated with the Eksa Formation (Boruchinkina *et al.*, 1974; *Geologicheskaya Karta* ..., 1981).

In the stratotype (Eksa Cliffs), the Eksa Formation underlies a basaltic flow, which is separated by a member of psammitic, aleuritic, and pelitic tuffs and tuffites from another flow occurring upstream the Nizhnyaya Tunguska River. The total thickness of this tuffolava unit, referred by Belozerov to the lower part of the Kholokit Group, is about 30 m. We consider it as the Khenochangda Formation, whose stratotype is distinguished along both sides of the Nizhnyaya Tunguska River between the Yambukan and Khenochangda River mouths. The stratotype includes the following bed succession:

1. Basalt, dark grey, fine-grained. Thickness: 1 - 6 m.

2. Aleuritic tuff, deep greenish-grey and black. Thickness: 6 - 12 m.

3. Tuff, deep greenish-grey, medium-grained. Thickness: 0 - 1.5 m.

4. Basalt, dark grey, fine-grained, massive at the base, with columnar jointing above, first vertical, then tilted. Thickness: 8 - 12 m.

The sequence is covered with variegated psephitic tuffs and tuffites with a total thickness of about 100 m. In our opinion, they represent the separate Kholokit Formation; however, the Khenochangda basalts should be excluded from the volume that was originally suggested by Belozerov, as the basalts are unexposed in the stratotype.

In the inner parts of the southwestern syneclise limb, the top sequence of the Korvunchana Group includes mainly aleuritic tuffosedimentary and volcanoclastic rocks, about 40 m thick. According to Belozerov, these rocks compose the Irgakta Beds (*Geologicheskaya Karta* ..., 1985; *Resheniya* ..., 1981), but we suggest that they correspond to the formation rank. The stratotype near the Irgakta Island of the Nizhnyaya Tunguska River, 8 - 10 km downstream from the Nidym mouth, reveals the following succession:

1. Aleuritic tuff, deep greenish-grey and black, fine-grained, enclosing interlayers of thin-platty varieties with small lensing concretions of limy siltstones and limestones bearing abundant conchostracan *Falsisca zavjalovi* (Nov.), *F. turaica* (Nov.), *Lioestheria evenkiensis* (Lutk.), *Limnadia vana* (Nov.), and some ostracode forms. Thickness: 0 - 1 m.

2. Aleuritic tuff, grey and deep greenish-grey, thin-platty, fine- to coarse-grained, with lensing interlayers of abundant concretions irregular in shape, which contain some conchostracan and numerous ostracode remains. Thickness: 4 m.

3. Tuffstone, deep greenish-grey (green when weathered), fine-grained, with lensing cross-bedding and fauna of ostracodes and poorly preserved conchostracan. Thickness: 2 - 3 m.

4. Gravelly vitroclactic tuffs, greenish-grey or brown, with black vitroclasts in a lighter colored matrix. Enclosed fragments of aleuritic tuffs and tuffites, sometimes abundant, add features typical of block breccia. The unexposed interval separates the base of the bed from the top. Suggested thickness: about 10 m.

5. Fine-grained tuffstone and coarse-grained aleuritic tuff, both grey- or brown-green, with lensing cross-bedding, thin gravelly interlayers, and very poorly preserved conchostracan. Thickness: 1 - 2 m.

6. Gravelly tuff enclosing grey pumice fragments cemented by a cherry-red matrix. Thickness: 3m.

7. Basalts (?), light grey, maculose, with autometamorphic (?) alterations at the base (0.5 - 2 m); the rock grades upward into the altered red tuffaceous variety. Thickness: 7 m.

8. Silty tuffite, grey-pink, maculose, with pisoliths and abundant root remains. Near the top, the rock grades into light grey, fine-grained aleuritic tuff. Thickness: 4 m.

Thus, in the inner parts of the southern syneclise limb, the Korvunchana Group consists of 8 units. It underlies lavas represented in the basal part by thin flows of amygdaloidal basalts, which were originally referred to the Tura Formation (Shorokhov, 1974a) and later renamed as the Nidym Formation (Boruchinkina *et al.*, 1974; Sadovnikov, 1974; *Geologicheskaya Karta* ..., 1985). Above is the Agitkan Formation (Sadovnikov, 1981; or the Chuvakan Formation, according to Boruchinkina *et al.*, 1974) of tuffites and tuffosedimentary rocks, which underlie the Kochechumo Formation of thick, poorly mineralized basaltic flows.

A well-known exposure near the Gagarii Ostrov (an islet) is 20 km downstream from where the Taimura flows into the Nizhnyaya Tunguska River. The exposure is disturbed by a small ravine. Upstream from the ravine, the stratotype of the Gagarii Ostrov Formation (or Horizon, according to Malovetskaya *et al.*, 1976) comprises light grey, fine-grained polymictic sandstones and dark grey siltstones underlying greenish-grey aleuritic tuffs with abundant remains of cordaites and ferns. Downstream from the ravine, the exposed rocks include lower grey and upper green or red aleuritic tuffs referred to as the Tutonchana Formation (Belozerov and Sadovnikov, 1962). Many geologists still believe the unit to be valid, and some even suggest that it be considered as the lectostratotype of the formation; this, however, is misleading. Originally correlating the above deposits with the Tutonchana Formation, one of the authors believed that they are older than the Eksa Beds, because their distribution area is more remote from the syneclise axial zone. In addition, the deposits were considered to be overlying the Gagarii Ostrov Horizon, which, at that time, also approved their correlation with the Tutonchana rocks. Both arguments are unreliable, and new observations contradict them.

First, large pebbles (up to 6 cm in diameter) from the base of volcanogenic beds of the Gagarii Ostrov stratotype consist of grey, gravelly tuffs. This suggests that, below, there are (or were, but now eroded) tuff beds. Second, extending downstream the Nizhnyaya Tunguska River, the exposed beds rise upward toward the northeast, and 3 km down, near the Iryakta Island, there is another exposure of tuffosedimentary beds (about 20 m) dipping toward the southwest. The grey, gravelly tuffs below them (about 15 m) are very similar to the rocks composing the pebbles at the base of the Gagarii Ostrov section. The tuffs, in turn, overlie another member of tuffosedimentary deposits (25 m). All of this suggests that the similar rocks in the Gagarii Ostrov section are not the basal in the Korvunchana Group. Moreover, the paleontological characteristics of aleuritic tuffs in the section also suggest their younger age as compared to the Tutonchana Formation.

In the outer zones of the southwestern limb of the syneclise, the Korvunchana Group reveals another succession. There are a few exposures of gravelly tuffs and xenotuffs, and they are all located along the Nizhnyaya Tunguska River between the Gagarii Ostrov locality and the mouth of the Chiskova River. Similar rocks are exposed in the valley of the Uchami River, 10 km upstream from the mouth. If we accept the diagnosis of the Uchami Formation that was suggested by Teben'kov for the standard stratigraphic scheme (Belozerov and Sadovnikov, 1962; *Geologicheskaya Karta* ..., 1981; *Resheniya* ..., 1981), then the section of the last locality should be considered as the stratotype. Note that older deposits assigned to the Uchami Formation by Shorokhov (1974b) were included into the Gagarii Ostrov Formation (Malovetskaya *et al.*, 1976). In the upper part of the Uchami valley, the gravelly tuffs contain peculiar, acute-angled clasts of various colors.

In the lower courses of the Korvunchana River, 3.2 km from its mouth, the exposed bituminous siltstones of the Gagarii Ostrov Formation are similar to rocks from the base of the Taimura River section. They overlie greenish-grey aleuritic tuffs alternating with silty or pelitic tuffs and tuffites, all considered by A.I. Emel'yanov and N.I. Emel'yanov as the basal subformation of the Korvunchana Formation. In 1957, these rocks were traced over the entire the Degali-Uchami interfluve area and ranked as the Tutonchana Formation (Sadovnikov, 1960). This formation underlies gravelly xenotuffs and is disturbed by a small hiatus in the middle part. Later, rocks below the hiatus yielded cordaites and were readdressed to the Gagarii Ostrov Formation. Gravelly tuffs overlying tuffosedimentary rocks occur in many outcrops of the Nizhnyaya Tunguska foreshore downstream from the Korvunchana mouth, where they are 100 - 150 m thick. Between the mouths of the Bol'shaya and Malaya Semenova rivers, the tuffs overlie the following sequence:

1. Sandstone (tuffaceous ?), grey, mottled, fine-grained. Thickness: 0.5 m.

2. Aleuritic tuff, greenish-grey, fine-grained, thin-platty, exposed after a covered interval (1 - 2 m). Thickness: 0.5 m.

The rocks below are exposed in the upper foreshore areas 2 m downstream. Above the aleuritic tuff there are:

3. Conglomerates with pebbles (1 - 10 cm) of light grey, coarse-grained aleuritic tuff, which are grey- or yellow-green on a weathered surface. In addition, the rock contains pebbles and angular fragments (0.2 m) of grey or brown-grey, fine-grained limestones, and large boulders (0.5 m) of gravelly tuffs. The cement consists of greenish-grey, fine- to coarse-grained silty material. Thickness: 2 m.

4. Aleuritic tuff, green to brown, fine- to coarse-grained, thick-platty, with tuffstone and conglomerate lenses. Thickness: 4 - 6 m.

5. Aleuritic tuffs, pink-grey, fine-grained with variegated clasts and remains of *Voltzia* (?). In the lower part, they enclose interlayers of greenish-grey, fine-grained rock varieties bearing *Madygenia, Lepidopteris* (?), and *Filices*. In the middle part, the interlayers consist of sandstones and gravelstones with variegated clasts; in the upper one, they are composed by cherry-red, fine-grained aleuritic tuffs. Total thickness: 3 m.

6. Aleuritic tuff, grey to light grey, fine-grained, with occasional pisoliths. At the base they contain rare *Madygenia* forms and abundant root remains. Thickness: 0.5 m.

7. Aleuritic tuff, pink-grey, fine-grained. Thickness: 2 m.

As the unequivocal correlation of these deposits is impossible, we consider them to compose the separate Semenova Formation, which is characterized by the above-described stratotype.

Between the Chiskova and Tutonchana rivers on the right side of the Nizhnyaya Tunguska River, there is a thick anamesite body elevated to a height of 400 m. In the lower Tutonchana course, 1 km above the Bugarikta mouth, this body overlies tuffs (about 200 m) with tuffosedimentary rocks (approximately 50 m) below them, both referred to the Bugarikta Formation (Belozerov and Sadovnikov, 1962). The anamesite occurs below gravelly tuffs (200 m) underlying the Agitkan and Kochechumo formations. Volcanogenic and sedimentary deposits similar to the Bugarikta rocks are exposed in the Nizhnyaya Tunguska foreshore, near the Bugarikta mouth. Here, in the lower foreshore, N.I. Emel'yanov and V.A. Fedotova distinguished grey terrigeneous deposits of the Gagarii Ostrov Formation (*Geologicheskaya Karta* ..., 1985). Thus, it is probable that in this area the Bugarikta Formation directly overlies the last formation, but they could also be separated by a tectonic boundary. Even if this contact is stratigraphic, the Bugarikta Formation is hardly correlative to the Tutonchana Formation; it underlies the anamesite body and, therefore, is younger than the Semenova Formation and the tuffs above it.

According to Boruchinkina *et al.* (1974) and Betekhtina and Mogucheva(1984), the basal tuffosedimentary deposits of the Korvunchana Group contain coniferous *Voltzia* (?) *avamica* Mog. in addition to Mesophytic forms of the fern and horsetail groups. Mogucheva also noted that they bear *Voltzia* (?) *avamica* Mog. in sections of the Detykte's middle course.

In general, the Korvunchana Group of the outer southwestern limb of the Tunguska syneclise (between the Chiskova and Bugarikta rivers) consists of three units. Its Semenova Formation, bearing *Madygenia* and *Voltzia* (?) forms, is correlative to the Dyuke Formation. Tuffs occurring between this and the Bugarikta Formation were mapped as rocks of the Uchami Formation. However, the noted similarity between the variegated Semenova gravelly tuffs and those overlying the stratotype section of the Uchami Formation suggests that the latter can be correlated with the Allyun Formation. Thus, the Bugarikta Formation corresponds to an interval of the Khenochangda and Irgakta formations, and the tuffs below it, to the Chomchrokon Formation.

In the syneclise axial zone between the Nidym and Kiramki mouths, the Korvunchana Group is unexposed. As follows from drilling results (Wells 7p and 11p between Babkino and Tura settlements), the thickness of the Chomchorokon tuffs is considerably decreased here (down to 0 - 30 m), in contrast to the thicker tuff beds of the Allyun Formation; other units are less variable in this respect.

The eastern limb area of the Tunguska syneclise, between the mouth of the Khovorkili River and Yastrebok Crossing, is defined as the central facies zone (*Resheniya* ..., 1981). Here, a series of well-correlative outcrops reveals a sequence of tuffosedimentary rocks (up to 60 m), variegated coarse tuffs (up to 60 m), basalt flows with tuff interlayers (up to 45 m), another tuffosedimentary member (up to 60 m), and gravelly tuffs (up to 50 m); all of them underlie basalts of the Nidym Formation. These rocks are distinctly correlative to respective rocks of the Irgakta and Chomchorokon formations distinguished in the axial syneclise region.

In the southern facies zone (*Resheniya* ..., 1981), the Korvunchana Group is subdivided into the Yuzhnaya Chunya (tuffosedimentary rocks up to 120 m thick), Chichikan (tuffs up to 430 m thick), and Limptekon (300-m-thick tuffs and sediments) formations (*Geologicheskaya Karta* ..., 1991; Orlova and Sadovnikov, 1974). In the Pirda River basin, the Limptekon deposits contain several (usually two) basaltic flows and can be referred to as the separate Pirda Formation, which is up to 180 m in thickness. In the upper course of the Pirda and adjacent areas, this formation underlies thick (up to 350 m) Igodekit tuffs with Kochechumo basalts above them. The Yuzhnaya Chunya Formation was earlier considered as correlative either to the Tutonchana Formation (*Geologicheskaya Karta* ...., 1981, 1985; *Resheniya* ..., 1981) or to the Dyuke and Eksa formations (*Geologicheskaya Karta* ..., 1991; Orlova and Sadovnikov, 1974). Presented data suggests that it corresponds better to the Dyuke (but not Eksa) and Semenova formations (the latter previously considered as the Tutonchana Formation).

It is now clear that all along the southern periphery of the Tunguska syneclise, the Korvunchana Group sequence begins with deposits corresponding to the lower unit of the Dvurogii Horizon. We suggested classifying these rocks as the Lebedev Horizon (Orlova, 1990, 1991; Sadovnikov, 1987). Their stratotype is in the Lebedev Creek, where the left tributary of the Dezertir Creek flows into the Avam River (Kureika River basin). The stratotype, which overlies basalts, anamesites, and aleuritic tuffs of the Verma Formation, includes the following beds:

1. Gravelly (sometimes psammitic) tuffite, greenish-grey, thin-platty. Thickness: from 1 - 2 m.

2. Gravelly tuff, greenish-grey, enclosing interlayers and lenses of gravel and ash tuffite. Thickness: 9 - 14 m.

3. Aleuritic tuffite, greenish-grey. Thickness: 1.5 - 2 m.

4. Gravelly tuff, greenish-grey, with tuffite interlayers. Thickness: 3 - 4.5 m.

5. Aleuritic tuffite, greenish-grey, fine-grained. Thickness: 0.5 - 1 m.

6. Gravelly tuff (xenotuff ?), greenish-grey, with dark and light vitroclasts. The bed encloses (9 m above the base) a 3-m-thick aleuritic interlayer composed of volcanoclastic laminae of variable grain size. The basal part of the interlayer (0.5 m above its base) bears abundant *Madygenia borealis* Radcz., rare *Radicites* sp., and *Peltaspermum* sp. in association with occasional *Voltzia* (?) cf. *avamica* Mog. Specimens of the latter are more abundant 1.5 m higher in the interlayer, where rare *Parajacutiella* (?) *avamica* Mog. is also present. Thickness: 26 - 32 m.

7. Aleuritic tuff, green- to yellow-grey, with interlayers of tuffite of variable (ash to gravel) grain size. Thickness: 13 m.

8. Gravelly tuff, greenish-grey. Thickness: 3.5 - 7 m.

9. Alternating interlayers of aleuritic, psammitic, and gravelly tuffites. Total thickness is 68 m. In the Dezertir Creek section, there are three fossiliferous horizons of this bed: the first (42 m above the base) contains abundant *Voltzia* (?) *avamica* Mog. and rare *Cladophlebis* cf, *kirjamkensis* Pryn.; the second (46 m above the base) contains abundant *Cladophlebis* sp., rare *Sphenopteris* typ. *trisecta* Schved., and single *Pecopteris* sp.; and the third (48 m above the base) contains occasional specimens of *Rhipidopsis* cf. *lobata* Halle. In the Lebedev Creek locality, the bed contains (8 m above the base) a lens (0.1 - 0.8 m thick) of yellow-brown, fine-grained aleuritic tuff with abundant *Elatocladus pachyphyllus* Pryn., rare *Cladophlebis* sp., and single *Pecopteris* (vel *Acrostichides*) typ. *polymorphum* Halle. The upper part of the lens bears abundant *Carpolithes* aff. *zwetkoviensis* Mog., cf. *Boweria* (vel *Acrostichides*) sp., and *Quadrocladus* (?) *sibiricus* (Neub.), along with rare *Acrostichides* (?) sp., *Todites* (?) sp., cf. *Mertensides* sp., and single *Glossophyllum* sp. and *Madygenia* sp. Rare specimens of *Neokoretrophyllites annularioides* Radcz., *Katasiopteris lebedevii* Radcz. f. *lata* (Mog.), and *Rhizomopteris* sp. occur 1m above the bed's base in outcrops of the upper left tributary of the Dezertir Creek.

10. Gravelly tuff, greenish-grey. Thickness: 4 m.

11. Aleuritic tuff, light green-grey, fine-grained, with ash tuffite interlayers. Total thickness: 16 - 23 m. In the Dezertir locality, the bed contains (3 m above the base) an interlayer (0.2 m) of pisolitic tuffite. The interlayer is psammitic at the base, where there are abundant remains of *Voltzia* (?) *avamica* Mog., *V. angusta* Mog., and *Pecopteris* aff. *zauronica* Pryn. Its upper aleuritic part bears abundant *Sphenopteris* typ. *microphylla* Neub. in col. and single specimens of *Voltzia* (?) *angusta* Mog. and cf. *Cladophlebis* (?) *sewardii* Johanss.

Along the left tributary of the Dezertir Creek, the sequence continues upward with the following beds:

12. Tuffite, light grey, with coarse gravel clasts at the base and aleuritic above it. The yellow aleuritic interlayer (5 m) near the top contains abundant *Madygenia borealis* Radcz. and single specimens of *Rhizomopteris* sp. Thickness: 8 cm.

13. Aleuritic tuff, grey to green-grey, platty and bedded due to the alternation of fine-grained and coarser varieties. Thickness: 10 m.

14. Aleuritic tuff, black, conchoidal. Thickness: 3 m.

15. Aleuritic tuff with clasts of variable grain size, grey to green-grey, platty. Thickness: 18 m.

According to Mogucheva (1973), the Dvurogii Formation of the syneclise's western limb (the Lebedev Horizon in our stratification) overlies deposits with a fossil plant assemblage similar to that of the Tutonchana Formation. If this is correct, we may specify these deposits as the Tutonchana Horizon. However, the above descriptions show that in the axial syneclise zone this horizon is distinguishable only in well sections. Other deposits of the zone (the Semenova and Yuzhnaya Chunya formations) do not belong to this horizon, though earlier they were assigned to the Tutonchana Formation. If it were possible to approve the correlation between the Semenova deposits and the Tutonchana Formation stratotype, then we could join the Tutonchana and Lebedev horizons indivisibly and specify this united subdivision by its prior name: the Korvunchana Horizon (Malovetskaya *et al.*, 1976; Sadovnikov, 1979).

Because the Ayan Formation above the Dvurogii deposits corresponds to the Nidym Formation (Sadovnikov, 1981), the Eksa and Irgakta formations should be placed at the level of the upper Dvurogii beds. In their floral assemblages, these units do not differ from overlying deposits and have been respectively included in the Putorana Horizon (Sadovnikov, 1987). Nevertheless, the conchostracan assemblages from the Eksa and Pirda formations are incompatible with those in the Putorana deposits (Orlova, 1991). This distinguishes the relevant host rocks as the Khungtukun Horizon, with its stratotype located between the Anakit and Khungtukun mouths on the right side of the Nizhnyaya Tunguska River. Here, in 1936, Shorokhov and colleagues first discovered conchostracan remains, which were later described by Lyutkevich. On the right side of the Nizhnyaya Tunguska River, 4.5 - 5.3 km downstream from the Anakit mouth, exposed beds represent the following succession:

1. Taxite basalt, green-grey, fine-grained. Thickness: 10 m.

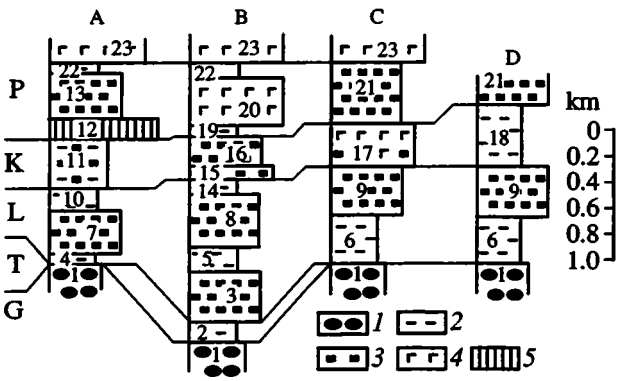

Fig. 1. Correlated sections of the Permian–Triassic volcanogenic deposits in the central Tunguska syneclise.

1 – sandstones, siltstones, and shales; *2* – ash tuffs, tuffites, and tuffosedimentary rocks; *3* – gravelly tuffs and xenotuffs; *4* – basalts; *5* – anamesites.

Regions: A – southwestern, the Korvunchana and Kochumdek rivers; B – central, the Babkino and Tura settlements; C – eastern, the Pirda River; D – southeastern, the Eika and Ilimpeya rivers. Horizons: P – Pirda; K – Khungtukun; L – Lebedev; G – Gagarii Ostrov. Formations marked by numbers: 1 – Gagarii Ostrov; 2 – Tutonchana (?); 3 – Allyun; 4 – Semenova ("Tutonchana"); 5 – Dyuke; 6 – Yuzhnaya Chunya; 7 – "Uchami"; 8 – Chomchorokon; 9 – Chichikan; 10 - 13 – Bugarikta; 14 – Eksa; 15 – Khenochangda; 16 – Kholokit; 17 – Pirda; 18 – Limptekon; 19 – Irgakta; 20 – Nidym; 21 – Igodekit; 22 – Agitkan (Chuvakan); 23 – Kochechumo.

The flow dips downstream toward a locality situated 6.3 km below the Anakit mouth where, presumably above the basalt, a bed consists of:

2. Tuffite, greenish-grey. Thickness: 2 m.

One and a half kilometers downstream from this bed (5 km above the Khungtukun mouth), there is another, presumably higher bed of:

3. Siltstone, pea-yellow, coarse- to fine-grained, platty, with one thin-platty interlayer bearing poorly preserved conchostracan *Falsisca* sp. and *Limnadia* sp. Thickness: 1 m.

After a covered interval of about 4 m, an outcrop in the upper foreshore displays:

4. Aleuritic tuffite, pea-yellow. Thickness: 5 m.

5. Taxite basalt, dark grey, fine-grained. Thickness: 10 m.

Upstream from the outcrop, the thickness of the basalt first decreases to 3 m and then increases to 20 m, i.e., it increases to the total height of the foreshore. Downstream from it, the basalt bed lenses down to a thickness of 3 m.

Because all these outcrops are poorly exposed, it is not inconceivable that Bed 3 is a lower one, if beds 1 and 5 represent a single basalt flow with tuffite (Bed 2) occurring above and below it. In the next outcrop, 3 km upstream from the Khungtukun mouth, this flow is below the following bed:

6. Aleuritic tuff, pea-yellow, fine-grained. Thickness: 3 m. At a level 2 m above the base, the bed contains an interlayer of brown limy siltstone with the following conchostracan species: *Limnadia vana* Nov. and *L.* aff. *vana* Nov. (abundant); *Lioestheria evenkiensis* (Lutk.) (rare); and *Eulimnadia* sp., *Falsisca* aff. *zavjalovi* (Nov.), and *Rohdendorfium* (*Bipemphigus*) *gennisi* (Nov.) (few). Calcareous concretions inside the bed also contain abundant *Limnadia vana* Nov. and less abundant *Echinolimnadia mattoxi* Nov. in association with rare *L.* aff. *vana* Nov., *Lioestheria evenkiensis* (Lutk.), *Eulimnadia* sp., and sporadic *Rohdendorfium* (*Bipemphigus*) *gennisi* (Nov.). In a nearby outcrop, the bed is exposed in the foreshore and basal part of the main steep slope. Here, its conchostracan assemblage includes abundant *Limnadia vana* Nov. and common *L.* aff. *vana* Nov. accompanied by rare *Lioestheria evenkiensis* (Lutk.), *Ragozinia leonidi* Nov., *Cyclotunguzites gazimuri* Nov., and single *Echinolimnadia mattoxi* Nov., *Eulimnadia* aff. *shorokhovi* Nov., *Falsisca zavjalovi* (Nov.), and *Rohdendorfium* (*Bipemphigus*) *gennisi* (Nov.).

Thus, it is possible to recognize the Tutonchana, Lebedev, Khungtukun, and Putorana horizons among volcanogenic rocks of the Tunguska syneclise (Fig. 1). Within the syneclise axial part, the Tutonchana Horizon comprises basal tuffosedimentary deposits of the Korvunchana Group. These deposits contain only fossil plant remains, mainly of ferns (*Todites, Acrostichides, Cladophlebis,* and *Katasiopteris*), associated with less abundant horsetails (*Neokoretrophillites* and others) and Gymnospermae (*Tersiella, Yavorskyia*). The unit correlates with the East European Vyatka Horizon of the Tatarian Stage. The Lebedev Horizon consists of tuffs and tuffosedimentary rocks. In the syneclise axial part (the Allyun and Eksa formations), it overlies either the Tutonchana or the Gagarii Ostrov horizons. In the syneclise limbs, this horizon (the Dvurogii, Yuzhnaya Chunya, and Chichikan formations) rests on the Gagarii Ostrov and older deposits. Its floral assemblages mainly include species of *Madygenia, Voltzia* (?), and *Rhipidopsis* genera, though in some areas they are also rich in fern forms similar to those in the Tutonchana deposits. The Paleophytic palynospectra of these deposits are always similar in their peculiarities to pollen assemblages from the Gagarii Ostrov and older formations. Relevant conchostracan assemblages are usually represented by *Rohdendorfium* (*Bipemphigus*) *gennisi* (Nov.), thus being indistinguishable from the same fauna of the Gagarii Ostrov Formation (Orlova, 1990). This suggests that the Permian Lebedev Horizon corresponds to the hiatus interval preceding the Vetluga Horizon of the East European platform. The Khungtukun Horizon above it includes tuffs, tuffosedimentary rocks, and basaltic members of the syneclise axial part. It comprises the Khenochangda, Pirda, upper Dvurogii, and lower units of the Bugarikta and Limptekon formations. Plant remains are represented mainly by conifers (*Quadrocladus*) and ferns (*Mertensides, Boreopteris,*

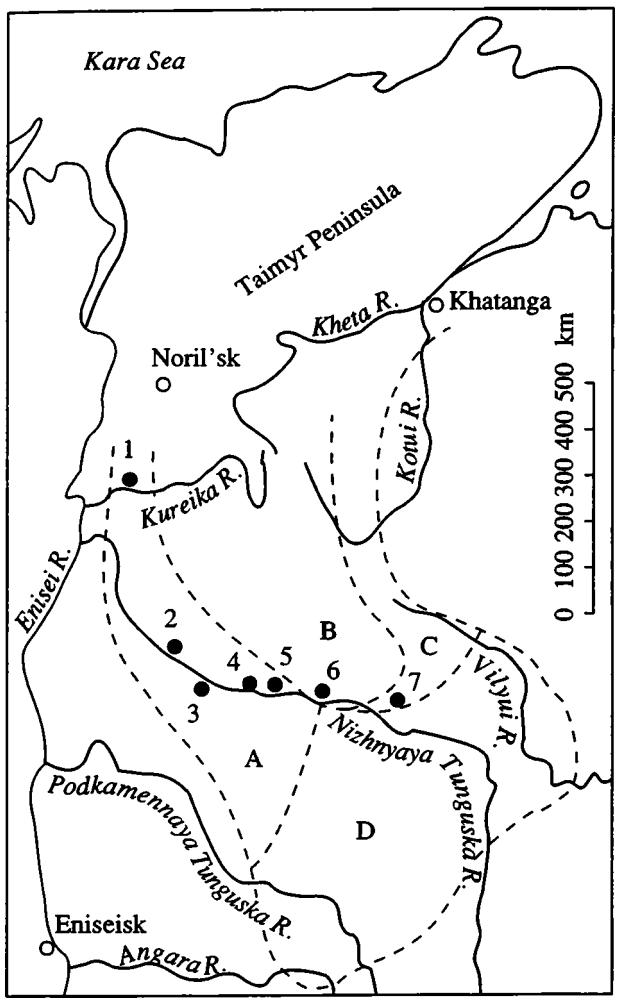

Fig. 2. Location scheme of main sections.
Letters for regions as in Fig. 1.
Sections: 1 – Avam; 2 – Bol'shaya and Malaya Semenova; 3 – Gagarii Ostrov; 4 – Eksa; 5 – Khenochengda; 6 – Irgakta; 7 – Khungtukun and Anakit.

Acrostichides, Cladophlebis), associated with less abundant horsetails and, in several localities, with *Lycopodium* forms (*Takhtajanodoxa, Tomiostrobus* (?). Predominantly Mesophytic spore and pollen assemblages often contain grains of *Lueckisporites*; but near the base, they are sometimes of the Paleophytic type. The conchostracan fauna is rich in species of *Leaioidea, Echinolimnadia, Lioestheria, Cyclotunguzites*, and *Limnadia*, accompanied by sporadic specimens of *Falsisca* and *Rohdendorfium* (*Bipemphigus*). The ostracode forms in different layers belong either to the Permian darvinulids or the genus *Gerdalia* (Orlova and Sadovnikov, 1974; Sadovnikov and Orlova, 1990). The Khungtukun Horizon also corresponds to a part of the hiatus preceding the East European Vetluga Horizon. The Putorana Horizon of the syneclise axial zone includes the lower Putorana basalts and the tuffosedimentary Irgakta Formation. In limb areas, it comprises tuffs of the upper Bugarikta and Igodekit formations, which are the stratigraphic counterparts of the lower Putorana Subformation (Nidym Formation) of the syneclise central zone. Floral assemblages of the Putorana Horizon are very similar to those in the Khungtukun Horizon. Its Mesophytic palynospectra often contain abundant grains of *Lueckisporites*. Conchostracan are almost always represented here only by *Falsisca* genus, and ostracodes, by *Gerdalia* forms. The age of the horizon is disputable, because its plant genera are typical of the Permian in contrast to the Triassic conchostracan and ostracodes (Sadovnikov, 1991).

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Phytostratigraphy and Correlation of the Jurassic Deposits of Western Siberia

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Abstract – The possibility of developing the phytostratigraphy of the terrestrial deposits of western Siberia without natural exposures is discussed. A detailed study of the taxonomic affinity of plant fossils, mainly groups that dominate the Jurassic paleoflora of the region, allowed for the specification of the taphofloras confined to definite levels of the stratigraphic sections. Changes in taphoflora composition were traced throughout the section. Significant stages in the Jurassic evolution of the taphofloras were revealed and used as the basis for distinguishing three regional phytohorizons of thick Jurassic deposits in western Siberia, which were further subdivided into beds with flora. These are the Urengoi (Lower Jurassic), Tomsk (Middle Jurassic), and Naunak (lower Middle–upper Upper Jurassic, Callovian–Oxfordian) horizons of defined regional distribution. The horizon ages are substantiated by the evolutionary level of the corresponding stratofloras and by rare finds of microfauna in the marine interbeds of the terrestrial Jurassic sections in the sublatitudinal Ob' and northern Tomsk regions. The Lower-Middle Jurassic boundary was placed at the base of the Tomsk Horizon, i.e., at the base of Member 2 (Radom) of the Kotukha Formation. The Tomsk and Naunak horizons were subdivided into beds with flora. Regional correlation of these beds revealed that the lower Jurassic section in the Tomsk region, the upper half of the Middle Jurassic section, and the lower part of the Vasyugan Formation are incomplete. The time limits of Jurassic sedimentation did not coincide in various areas of western Siberia.

The Jurassic deposits of western Siberia, the terrestrial part of which is traditionally recognized as the Tyumen' Formation (*Stratigraficheskii* ..., 1978), still remain poorly studied. This is especially true for the biostratigraphy of the Lower and Middle Jurassic deposits represented mainly by terrestrial sediments with an apparent polyfacial character and variations along the strike. In addition, these deposits, which are poorly studied by drilling, are less fossiliferous than the Upper Jurassic and Lower Cretaceous strata. Nevertheless, a great number of recent publications deal with the Lower-Middle Jurassic stratigraphy of western Siberia, particularly with lithological characteristics, cyclostratigraphy, reference horizons, geophysical properties, and palynological age datings of the deposits. Although they discuss sections recovered from the same wells, these publications suggest different approaches to subdividing the formation (Gugari *et al.*, 1988, 1992; Kazakov and Devyatov, 1990; Braduchan, 1985; Braduchan *et al.*, 1968; Nezhdanov and Ogibenin, 1987; Danenberg *et al.*, 1979; Rovnina *et al.*, 1978; Kabanova *et al.*, 1989; Egorova, 1989; and others) as well as several stratigraphic schemes for the Jurassic of western Siberia. Respectively, estimated ages of recognized units based on palynological data and rare foraminifers (Bulynnikova and Yasovich, 1972; Tat'yanin, 1979; Komissarenko, 1987, 1989; Tylkina and Komissarenko, 1977; Rovnina, 1976; Rovnina *et al.*, 1985; Purtova, 1989; Il'ina, 1985; and others) have been significantly corrected many times. These publications specified the geological structure of the Jurassic deposits, but they did not solve main biostratigraphical problems, i.e., the substantiation of regional units and detailed correlation of the deposits, especially those of terrestrial origin, which are still insufficiently characterized by plant megafossils. This is due to more than just the limited drilling of this part of the section, poor recovery of cores, and insufficient material for elaboration of the biostratigraphic framework. Until recently, the available collections of fossil plants from both the wells and the key sections in the western Siberian margins had not been adequately studied, and the taxonomic affinity of most plant remains had not been revised. This especially concerns the gymnosperms that dominate most western Siberian taphofloras. Moreover, available lists of plant fossils recovered from wells (Mogucheva, 1990) and different schemes for the formational division of terrestrial Jurassic deposits in the region often complicate the comparison and usage of paleofloral data.

In 1986, in order to elaborate the main principles of phytostratigraphic subdivision, correlation, and dating of terrestrial Jurassic deposits of western Siberia, we studied plant fossils from the stratotype and key sections of its southern and southeastern areas. A phytostratigraphic scheme was suggested for the Kuznetsk (Kirichkova *et al.*, 1992) and Kansk (Kirichkova and Travina, 1990) coal-bearing fields and the Tomsk region (Bystritskaya, 1992; Bystritskaya *et al.*, 1992). The refined biostratigraphic subdivision and correla-

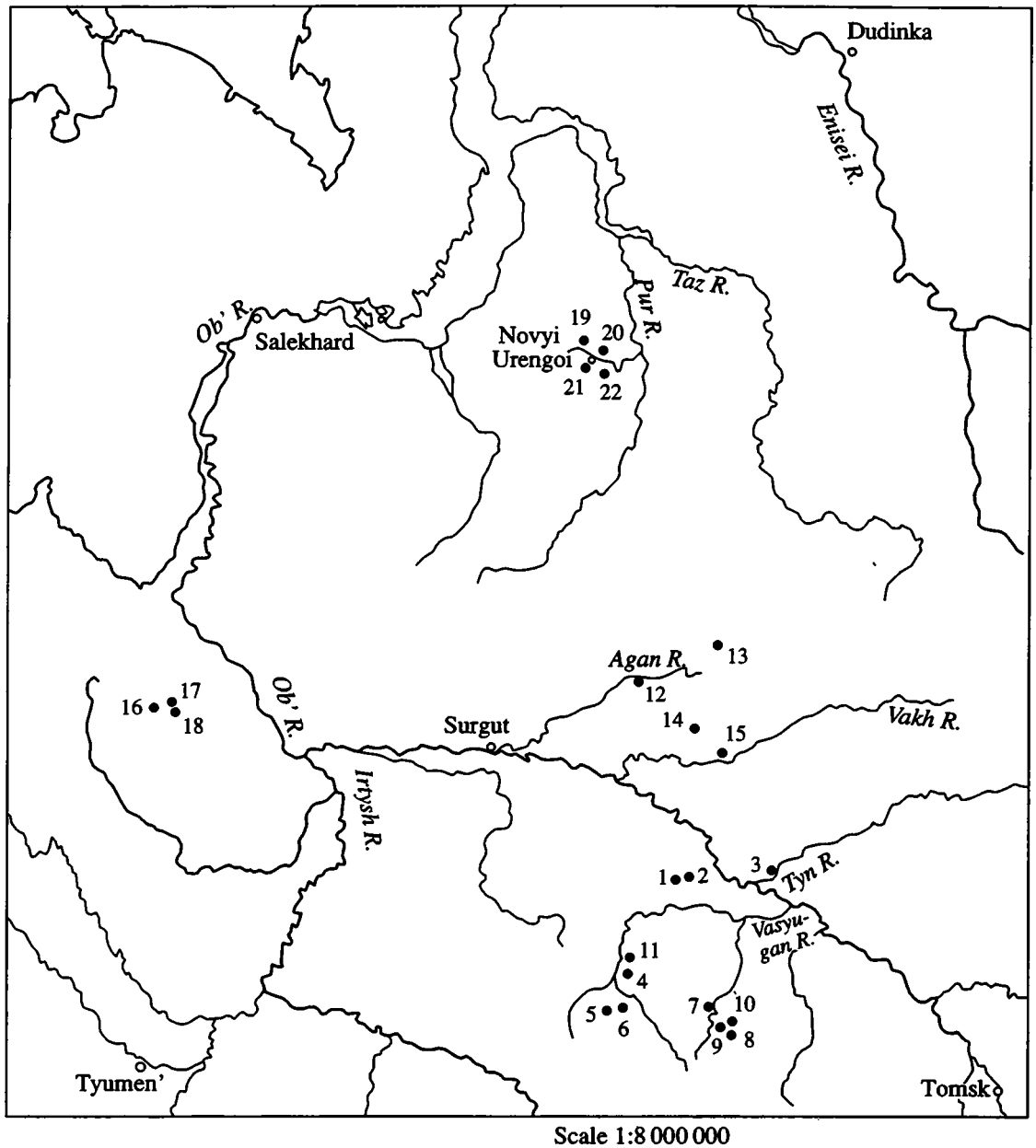


Fig. 1. Main well localities in western Siberia.

1 – Prikoltogorovo R-2; 2 – Southern Koltogorovo R-1; 3 – Tolparovo R-1; 4 – Southern Poselok R-1; 5 – Fedyushkino R-2; 6 – Northern Chertalinka R-400; 7 – Western Luginets R-180; 8 – Gerasimovka R-8; 9 – Southern Tambaevka R-76; 10 – Western Ostaninka R-447; 11 – Verkhnezayachii R-80; 12 – Var'-Egan p-952; 13 – Bakhil R-127; 14 – Southern Enitor R-165; 15 – Koshil R-304; 16 – Yakhlinka R-49; 17 – Talinka R-2797; 18 – Talinka R-129; 19 – Urengoi 673; 20 – Urengoi 411; 21 – Urengoi 414; 22 – Urengoi 410.

tion of the terrestrial Jurassic deposits of these areas are based on specific taphofloras with differentiated and stratigraphically significant taxa. An analysis of the original paleoflora collected from many wells in western Siberia and the taxonomic classification of typical western Siberian groups (Czekanowskiales, Ginkgoals, and some conifers) using epidermal-cuticular analysis (Samylina and Kirichkova, 1991; Kirichkova and Samylina, 1979, 1983; Kirichkova and Moskvin, 1989; Samylina and Markovich, 1991) provide an approach to refining the Jurassic phytostratigraphy of this territory for the first time.

The most complete sections of the terrestrial Jurassic of western Siberia were recovered from wells in the Urengoi area, in the most depressed area of the northern part of the western Siberian plate (Fig. 1). These coastal-marine and terrestrial Jurassic deposits are of maximal stratigraphic range and thickness. The scheme adopted at the interdepartmental stratigraphic meeting in Tyumen', 1990, envisages the subdivision of the

Table 1. Fossil plants from the lower part of the Jurassic deposits of the Urengoi area

Stratigraphic position (group, formation, well, depth, m) / Species	Tampei Group						Formations											
							Beregovoi			Yagel'		Kotukha						
	673				414, 5190 - 5258	410, 5415 - 5430	414, 5190 - 5258	410		673, 4974 - 4993	411, 5300 - 5317	673		414		411		410, 4671 - 4687
	5399 - 5417	5417 - 5469	5487 - 5503	5515 - 5526				5270 - 5278	5293 - 5300			4690 - 4709	4860 - 4870	4250 - 4267	4464 - 4498	4815 - 4830	4990 - 5005	5273 - 5290
1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20
Algites sp.																+			
Neocalamites cf. *carcinoides* Harris	+																		
Neocalamites sp.		+		+	+	+	+	+	+	+	+								
Paracalamites sp.		+	+																
Equisetites beanii (Bunb.) Sew.										+							+		
Equisetites sp.								+						+					
Phlebopteris polypodioides Brongn.												+				+			
Coniopteris cf. *latifolia* Brick														+					
Coniopteris sp.												+				+			
Todites cf. *princeps* Kacib.								+											
Cladophlebis sp.																	+		+
Sphenobaiera magnifolia Aksarin																			+
Czekanowskia ex. gr. *rigida* Heer														+					
Florina (?) sp.																+			+
Schizolepis cf. *moelleri* Heer														+					
Pityospermum cf. *nanseni* Nath.														+					

Lower Mesozoic of the Urengoi area into the following units (from the base upward): the Tampei Group (Upper Triassic), with the Varengaya and Vityutino formations, followed by the Beregovoi, Yagel', Kotukhta (Lower Jurassic), Tyumen' (Middle Jurassic), and Abalakovo (Callovian–Oxfordian) formations. The section from Well 414 (Figs. 1 and 2) was suggested as the stratotype of the Tampei Group and the Beregovoi Formation (Bochkarev *et al.*, 1989) and as the reference section of the Yagel' Formation.

The Tampei Group, penetrated by Well 414 to a depth of 5130 - 5287 m (157 m thick), lies unconformably on the Paleozoic deposits (Fig. 2). It comprises two units: the Varengaya Formation, represented by homogenous thin-bedded claystones with sandstone laminae and interbeds of conglomerates; and the conformably overlying Vityutino Formation, composed by sandstones with claystone and conglomerate members (depth range 5130 - 5188 m). The Tampei Group is referred to the Ladinian–Carnian according to palynological data (Bochkarev *et al.*, 1989). In the lower part of the group (Well 414), probably in the Varengaya Formation, we found the stem remains of horsetails *Neocalamites* (Table 1). Well 673 recovered the Tampei Group at a depth of 5330 - 5526 m. Here, the lower claystone interbed, probably corresponding to the Varengaya Formation, (depth range 5399 - 5526 m), contains stem remains of *Neocalamites* cf. *carcinoides* Harris (Table 1). Conchostracan remains were found at a depth of 5469 m in the same well. Wells 410 and 411 recovered only an upper part of the Tampei Group (the Vityutino Formation?), represented by sandstones with siltstone and conglomerate interbeds (Fig. 2) containing *Neocalamites* sp. The probable correlatives of the Tampei Group, which are only characterized by palynological data, were discovered in the Chernichnyi area in the north of western Siberia (Bochkarev *et al.*, 1989).

In the Urengoi area, the Tampei Group is overlain unconformably by the Beregovoi Formation. The hypostratotype of this formation is a section recovered from Well 414 at a depth of 4928 - 5130 m (Fig. 2). The formation consists of coarse-grained grey and greenish-grey sandstones with rare conglomerate and lenslike silty-clayey interbeds (Bochkarev *et al.*, 1989). Its time span is referred to the Hettangian–Sinemurian, which is the base of the Pleinsbachian according to the palynospectra. Plant fossils are scarce and represented by shoots of horsetails (*Neocalamites* sp.) found at a depth

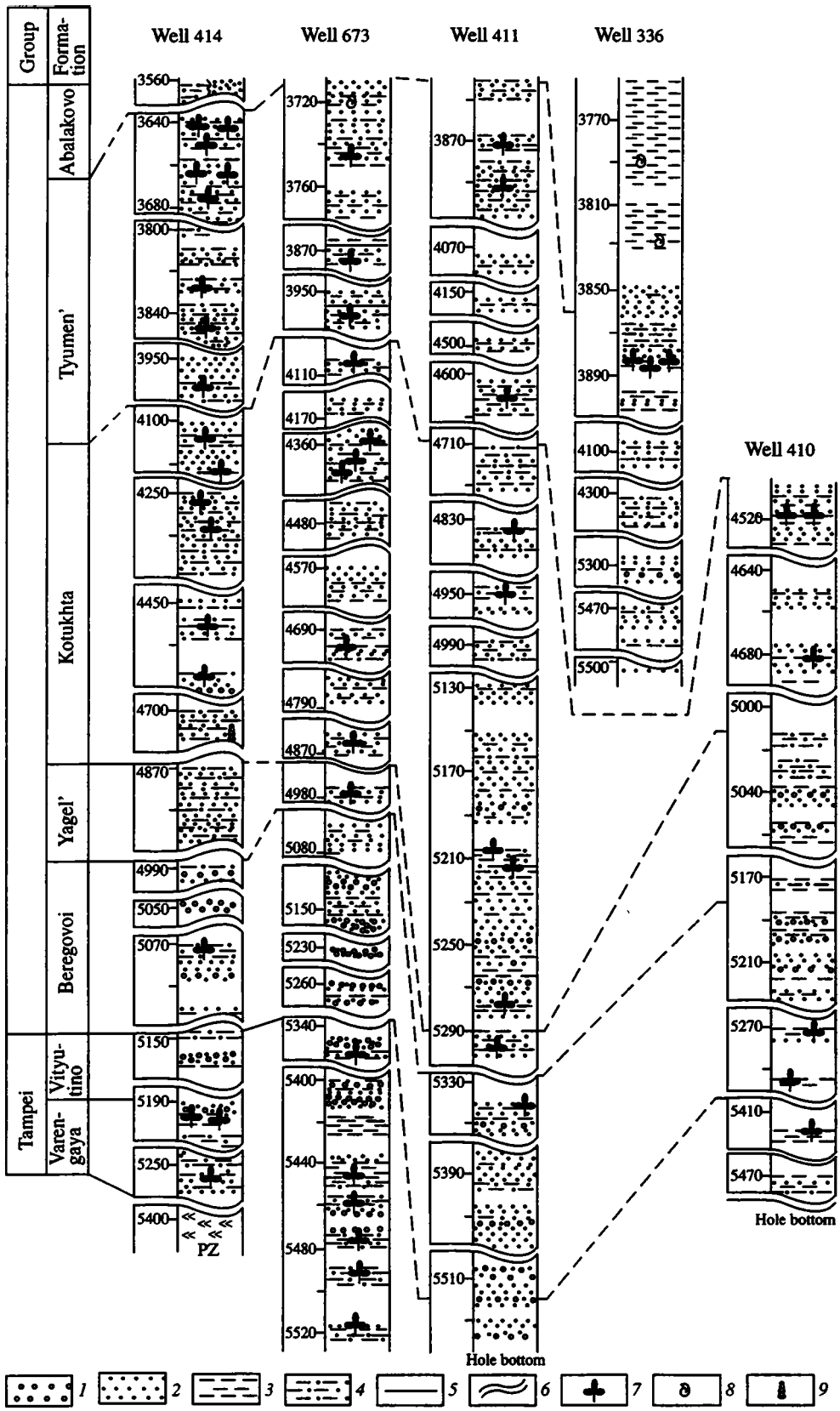


Fig. 2. Correlated key well sections in the Urengoi area according to paleofloral data.
1 – conglomerates; *2* – sandstones; *3* – claystones; *4* – siltstones; *5* – coal; *6* – no recovery; *7* – plant fossils; *8* – ammonite remains; *9* – foraminifers.

of 5075 m in Well 414. The same formation of coarse-grained sandstones lacking any plant fossils (Fig. 2) was recovered from Well 673 at a depth of 5060 - 5330 m. Remains of horsetail shoots (*Neocalamites* sp.) and a fragment of a pinnate leaf of *Todites* cf. *princeps* Racib. were found in the Beregovoi Formation, recovered from Well 410 at a depth of 5190 - 5347 m. The formation is the terrestrial conterpart of the Zima Formation in the Ust'-Enisei region, which bears forarminifers (Gurari *et al.*, 1988). No analogs of the Beregovoi Formation are currently known south of the Urengoi field.

The Yagel' Formation (in the scheme adopted by the Tyumen' meeting in 1990), or the Tarko-Sale Formation (according to the interpretation of SNIIGIMS geologists; Kazakov and Devyatov, 1990), was recovered from Well 414 at a depth of 4354 - 4928 m. The formation is composed of clays with thin interbeds of siltstones and fine-grained sandstones (Bochkarev *et al.*, 1989). It overlies the Beregovoi Formation and is poorly characterized by fossil shoots of *Neocalamites* sp. and *Equisetites beanii* (Bunb.) Sew., which were found only in Well 673 (Table 1; depth range 4974 - 4993 m). This formation probably corresponds to deposits recovered from wells 410 and 411 at depths of 5010 - 5180 m and 5290 - 5325 m, respectively. These are sandier deposits, with conglomerate interbeds in some areas (Fig. 2). Remains of shoots of *Neocalamites* sp. were recovered only from Well 411 at a depth of 5300 m.

The marine Levinskii Formation of north Siberia, which includes the upper Pliensbachian *Trochammina lapidosa* Beds (Komissarenko, 1987; *Stratigrafiya* ..., 1976), is an age analog of the Yagel' Formation (Table 2). Southward in western Siberia, there are no apparent correlatives of the Yagel' Formation. Pliensbachian deposits are recognized by palynological and geophysical records in some areas of the Nyurol'ka depression, Tomsk region, but their age is not supported by the assemblages of plant megafossils.

According to the scheme adopted in Tyumen', 1990, the Lower Jurassic deposits of the Urengoi field are completed by the Kotukhta Formation. The stratotype of this formation, represented by an alternation of clays, siltstones, and sandstones with gravelstone interbeds, is distinguished in the Bol'shaya Kotukhta area (Well 105-R, depth range 3552 - 3705 m; Nezhdanov and Ogibenin, 1987). In the Urengoi area, this formation, penetrated by several wells, contains plant fossils, although they are sparse (Fig. 2). In Well 673, this formation was recognized at a depth of 4690 - 4870 m. Its lower part yielded seeds of ancient Pinaceae *Pityospermum* and *Schizolepis* (Table 1); the middle part contains *Phlebopteris polypodioides* Brongn. and *Coniopteris* sp. (Fig. 2); and the upper one (possibly the analog of the Radom Member in the southern regions of western Siberia) bears *Equisetites* sp., *Nilssonia* cf. *kendali* Harris, *Sphenobaiera* ex gr. *czekanowskiana* (Heer) Flor., *Czekanowskia* ex gr. *rigida* Heer, and *Pityophyllum* sp. (depth range 4097 - 4489 m). Leaf remains of *Phlebopteris* and *Florinia* were found in the lower and middle parts of the Kotukhta Formation, wells 411 and 410 (Fig. 2, Table 1).

The lower age limit of the Kotukhta Formation in the Urengoi area is constrained by rare finds of foraminifers in the middle and lower parts of the formation. According to the resolutions of the Tyumen' meeting, the foraminifers were defined as corresponding to the Toarcian distribution range (Fig. 2). The Dzhangoda and Laida formations of northern Siberia, which contain Toarcian and early Aalenian fauna (Bulynnikova, 1972; Komissarenko, 1987) are the marine counterparts of the Kotukhta Formation (Table 2).

The fossil plant assemblages distinguished in the Kotukhta Formation are very indicative, although sparse. In the lower and middle parts of the formation recovered from wells in the Urengoi area, there are *Equisetites* cf. *beanii* (Bunb.) Sew., *Phlebopteris polypodioides* Brongn., *Sphenobaiera magnifolia* Aksarin, *Florinia* (?) sp., *Schizolepis* cf. *moelleri* Sew., and *Pityospermum* cf. *nanseni* Nath, which best correspond on an evolutionary level to the paleoflora of the second half of the Early-Lower Jurassic of the Siberian paleofloral province (Vakhrameev *et al.*, 1972; Vakhrameev, 1988). We found an analogous assemblage in the Sherkaly Formation (at the level of Bed Yu_{10}) of the Krasnoe Lenino arch (Well 129, the Talinka area, depth range 2667 - 2705 m). This assemblage includes *Todites princeps* (Presl.) Goth., *Lobifolia* sp., *Czekanowskia irkutensis* Samyl. et Kiritch., and *Ginkgo ananievii* Bystr.

A fossil plant assemblage from the upper part of the Kotukhta Formation in the Urengoi area seems younger, because the genus *Coniopteris*, represented by the species *C.* cf. *latifolia* Brick and *C.* aff. *vsevolodii E.* Leb), plays a more important role in its composition (Table 3).

The Kotukhta Formation is correlated with the terrestrial or coastal-terrestrial Khudosei Formation of the sublatitudinal Ob' and Tomsk regions (Nezhdanov and Ogibenin, 1987). The upper part of the latter, recovered from Well 962 in the Var'-Egan area at a depth of 3220 - 3420 m, produced an assemblage including *Equisetites lateralis* (Phill.) Phill., *Coniopteris* cf. *maakiana* (Heer) Pryn., and *Coniopteris* sp. This assemblage is similar to that from the upper part of the Kotukhta Formation in the Urengoi area.

In the Tomsk region, the Khudosei Formation is widespread and well-recovered; but in many wells it has an incomplete stratigraphic range (Fig. 3). As a rule, the upper part of the formation was recovered, which is represented by alternating sandstones and siltstones; locally, the predominant rocks are coarse- and medium-grained sandstones, black claystones, and carbonaceous siltstones. We identified the taphoflora from this part of the section as Tomsk assemblage 1 (Table 3) of the Tyumen' Formation (Bystritskaya *et al.*, 1992). In the Tomsk region, the upper part of the Khudosei

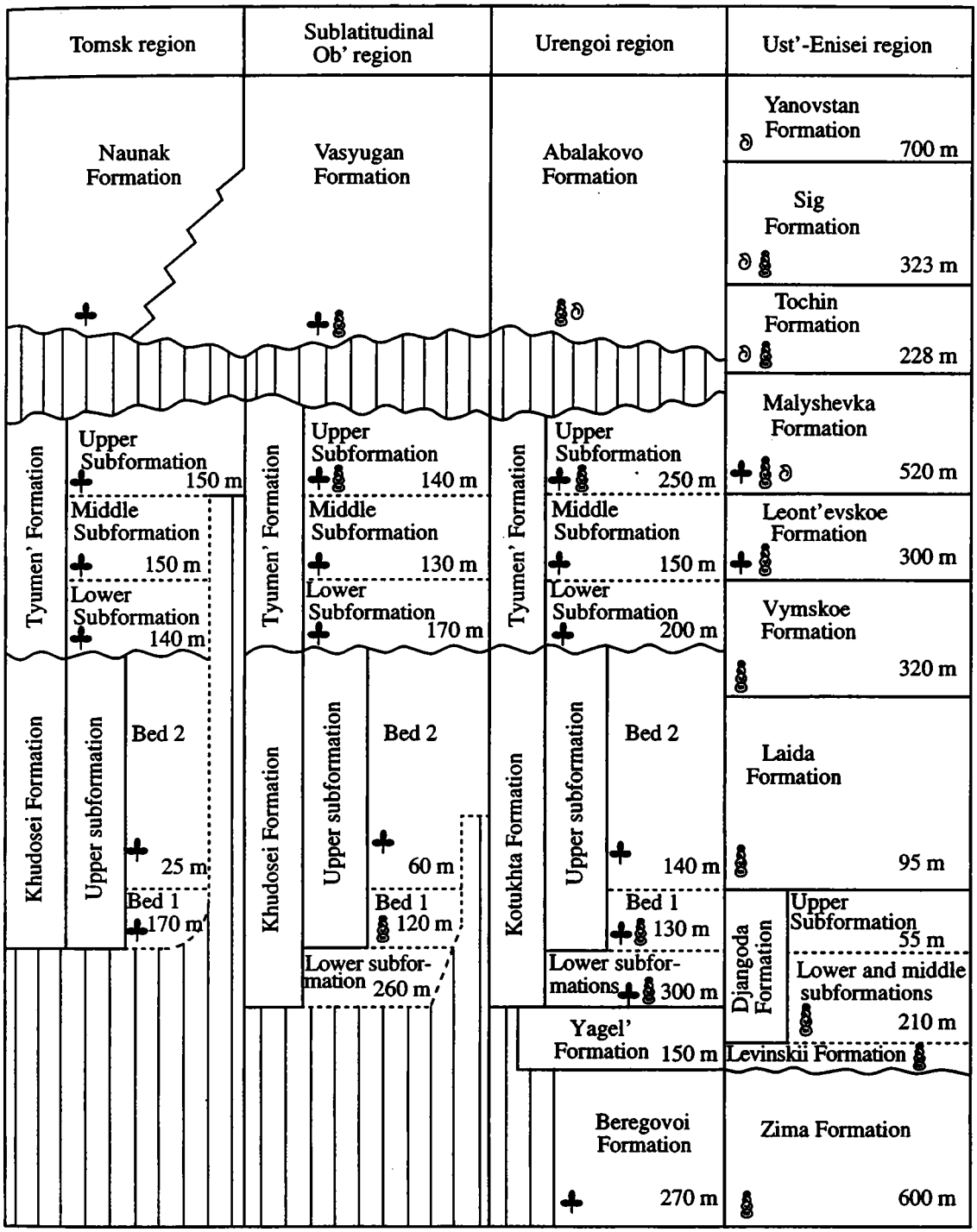
Table 2. Correlation scheme for the Jurassic deposits of western Siberia

Series	Stage	Horizon	Foraminiferal beds	Floral beds	Plant assemblages
Upper	Kimmeridgian	Naunak			
	Oxfordian		*Recurvoides disputabilis*	Beds with *Coniopteris latilobus, Czekanowskia tomskiensis, Phoenicopsis sibirica*	*Coniopteris latilobus* Bystr., *C. vialovae* Tur.-Ket., *Nilssonia urmanica* Bystr., *Czekanowskia tomskiensis* Kiritch. et Samyl., *Cz. vera* Kiritch. et Samyl., *Phoenicopsis sibirica* Kiritch. et Trav.
			Recurvoides scherkalyensis, Globulina paalzovi		
Middle	Callovian				
	Bathonian	Tomsk	*Recurvoides anabarensis, Globulina praecircumplia*	Beds with *Equisetites lateralis, Coniopteris depensis, C. simplex*	*Equisetites lateralis* (Phill.) Phill., *Coniopteris depensis* E. Leb., *C. simplex* (*L.* et *H.*) Harris, *Czekanowskia jenisseiensis* Kiritch. et Samyl.
			Trochammina praequamata		*Coniopteris hymenophylloides* (Brong.) Brong., *Raphaelia diamensis* Sew., *Nilssonia baranovae* Burak., *Phoenicopsis samylinae* Kiritch. et Moskv.
	Bajocian				
	Aalean			Beds with *Coniopteris maakiana, Leptotoma sibirica, Phoenicopsis irkutensis*	*Coniopteris maakiana* (Heer) Pryn., *C.* cf. *embensis* Brick, *Cladophlebis haiburnensis* (*L.* et *H.*) Sew., *Leptotoma sibirica* Kiritch. et Bat., *Ginkgo ananievii* Bystr., *Czekanowskia irkutensis* Kiritch. et Samyl., *Phoenicopsis irkutensis* Dolud. et Rasskaz.
Lower	Toarcian	Urengoi	*Riyadhella syndascoensis*		*Neocalamites* sp., *Coniopteris* sp., *Phlebopteris polypodioides* Brongn., *Todites princeps* (Presl.) Goth., *Pityospermum nanseni* Nath., *Florinia* sp.
			Saccammina inanis, Amoboculus lobus		
	Pliensbachian		*Trochammina lapidosa*		*Neocalamites* sp., *Cladophlebis* sp.
			Textularia areoplecta		

Formation is best represented in sections of Well 1-R of the Prikoltogorovo area, Well 1-R of the Tolparovo area, Well 180 of the Zapadnyi Luginets area, and Well 8 of the Gerasimovka area (Fig. 3). It lies unconformably on the Paleozoic deposits or on weathering crusts in the Zapadnyi Luginets and Gerasimovka areas.

In the Urengoi area, the Kotukhta Formation is overlain with an unconformity by the Tyumen' Formation (Fig. 2). Because of the deficiency of core material, the latter is poorly characterized by plant fossils throughout western Siberia, including the Urengoi area. The formation is generally represented by alternating siltstones, clays, and sandstones, often bearing

Table 2. (Contd.)



a detritus admixture (*Stratigraficheskii* ..., 1978). In the adopted scheme, the Tyumen' Formation is subdivided into lower, middle, and upper subformations. Abundant stem remains of *Equisetites* sp. and leaves of *Raphaelia* and rare specimens of *Coniopteris* and *Czekanowskia* were found in the lower part of the formation (the lower and middle subformations?), recovered by wells in the Urengoi area (Table 1, Fig. 2). The upper part of the Tumen' Formation, probably corresponding to the upper subformation, yielded a more diverse assemblage with frequent *Coniopteris* and abundant *Equisetites lateralis* (Phill.) Phill. (Table 1).

The marine and coastal-marine Vymoskoe, Leont'evskoe, and Malyshevka formations of the

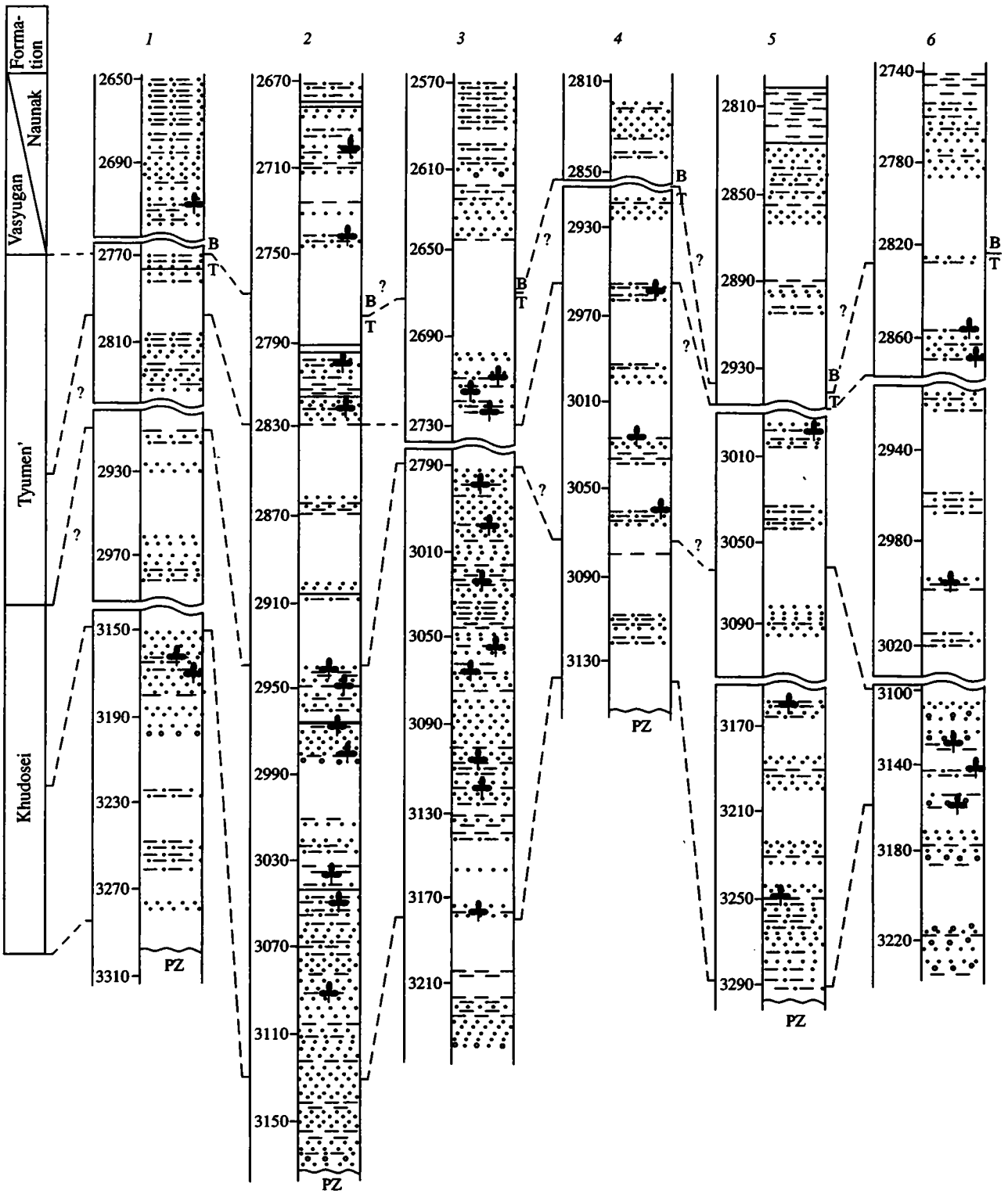


Fig. 3. Correlated well sections in the sublatitudinal Ob' and Tomsk regions according to paleofloral data.
Wells: *1* – Prikoltogorovo R-2; *2* – Southern Koltogorovo R-1; *3* – Tolparovo R-1; *4* – Southern Poselok R-1; *5* – Fedyushkino R-2; *6* – Northern Chertalinka R-400; *7* – Western Luginets R-180; *8* – Gerasimovka R-8; *9* – Southern Tambaevka R-76; *10* – Var'-Egan p-952; $\frac{B}{T}$ – boundaries between formations according to geophysical data (For symbols see Fig. 2).

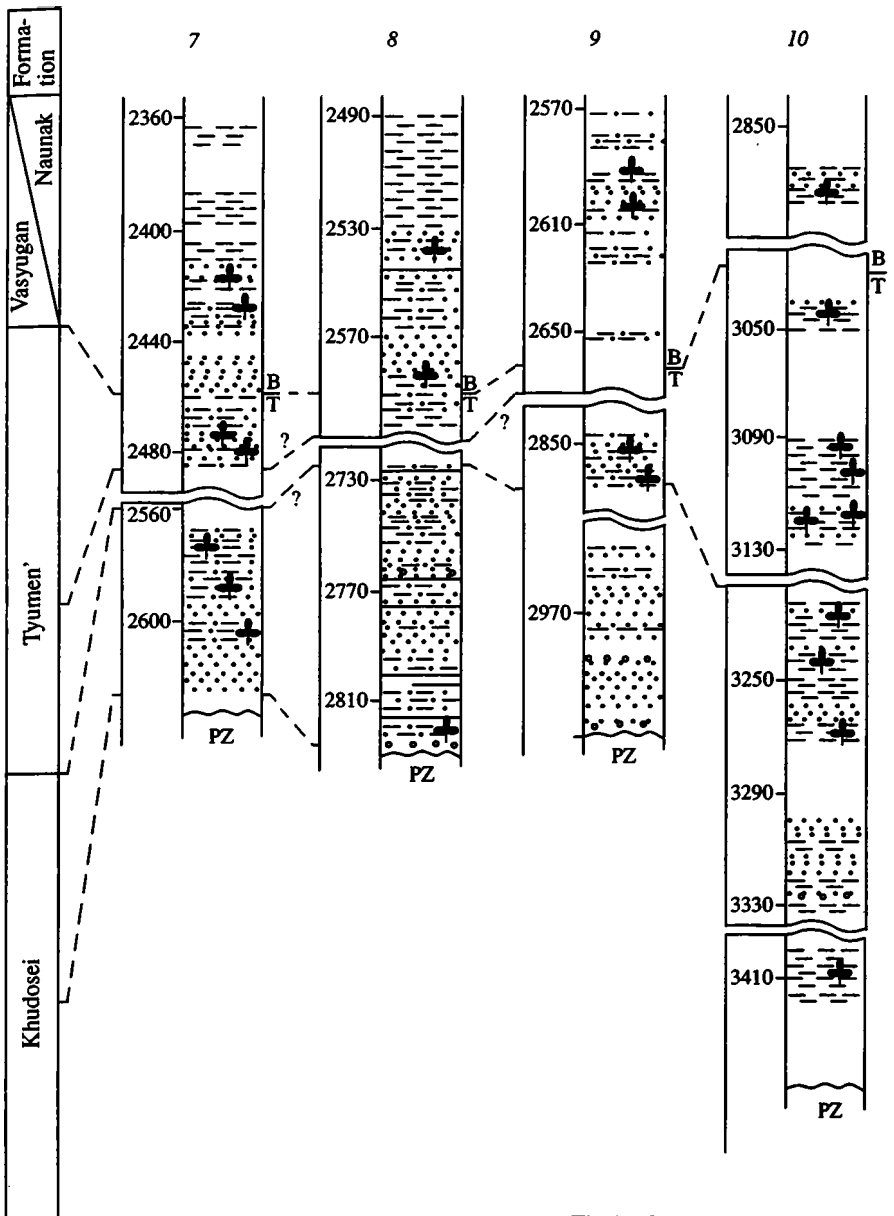


Fig 3. (Contd.)

northern region of western Siberia, containing Bajocian and Bathonian foraminifers (Bulynnikova and Yasovich, 1972; Komissarenko, 1987), are the analogs of the Tyumen' Formation of the Urengoi area (Table 2). The Malyshevka Formation is characterized by plant fossils as well (Mogucheva, 1990), but the published lists of flora disregard the oil fields and cannot be used for biostratigraphical purposes.

The Tyumen' Formation unconformably overlies the Sherkaly Formation of the Krasnoe Lenino arch and the Khudosei Formation of the Tomsk region. The Tyumen' Formation is composed of alternating grey sandstone, siltstone, and claystone interbeds with rare and thin coal laminae and conglomerate interlayers. In the adopted scheme, the formation is also subdivided into lower, middle, and upper subformations (Table 7). However, the core recovery from the lower and middle subformations is not sufficient (Fig. 3). Thus, they are poorly characterized by plant remains (Table 4).

The plant assemblage found in the Krasnoe Lenino includes *Coniopteris* cf. *simplex* (*L.* et *H.*) Harris, *Nilssonia* cf. *gracillima* Pryn., *Czekanowskia* cf. *jenissejensis* Kiritch. et Samyl., *Cz. rigida* Heer, *Cz. ferganensis* Kiritch. et Samyl., *Phoenicopsis* aff. *angustifolia* Heer, *Pseudotorellia abaniensis* Kiritch., and *Elatocladus* sp., yielded from Well 49 in the Yakhlinka area (depth range 2347 - 2445 m). Remains of *Equisetites beanii* (Bunb.) Sew., *Coniopteris* sp., *Czekanowskia* cf. *jenissejensis* Kiritch. et Samyl., *Cz. kanensis* Kiritch. et Samyl., *Cz. rigida* Heer, *Phoenicopsis dentata* Pryn., *Leptotoma* sp., and *Ginkgo* sp. were found in Well 2797 of the Talinka area (depth range 2537 - 2586 m).

Table 3. Plant assemblage from the upper part of the Khudosei and Kotukhta formations

Species \ Locality, well, depth, m	Tomsk region							Sublatitudinal Ob' region, Var'-Egan, 952, 3220 - 3420	Urengoi region				
										673			
	Gerasimovka, 8, 2815 - 2825	W. Luginets, 180, 2560 - 2610	S.-Koltogorovo, 1-R, 2944 - 3110	Tolparovo, 1-R, 2972 - 3179	Prikoltogorovo, 2-R, 3160 - 3168	N.-Chertalinka, 400, 3120 - 3160	Fedyushkino, 2-R, 3260 - 3280		414, 4106 - 4267	4097 - 4113	4350 - 4367	4470 - 4489	410, 4541 - 4556
1	2	3	4	5	6	7	8	9	10	11	12	13	14
Aligithes sp.								+					
Equisetites lateralis (Phill.) Phill.		+		+				+					
E. cf. *beanii* (Bunb.) Sew.													+
Equisetites sp.									+	+	+		
Coniopteris maakiana (Heer) Pryn.	+	+		+			+	sp.	sp.			sp.	
C. cf. *embensis* Pryn.				+									
C. hymenophylloides (Brongn.) Sew.		+						+					
C. cf. *latifolia* Brick									+				
C. cf. *nerifolia* Genk.							+						
Cladophlebis haiburnensis (L. et H.) Sew.	+												
Raphaelia sp.			+										
Nilssonia cf. *kendalli* Harris											+		
Leptotoma sibirica Kiritch. et Bat.	+	+											
Sphenobaiera ex. gr. *czekanowskiana* (Heer) Flor.	sp.				sp.							+	
Ginkgo aff. *sibirica* Heer					+								
G. cf. *ananievii* Bystr.				+		+							
G. ex. gr. *sibirica* Heer	+												
Czekanowskia rigida Heer	+	cf.								ex. gr.	ex. gr.		
Cz. baicalica Kiritch. et Samyl.						+							
Cz. jenissejensis Kiritch. et Samyl.			+	+		+		+					
Cz. cf. *ferganensis* Kiritch. et Samyl.			+										
Cz. obiensis Kiritch. et Samyl.					+								
Cz. irkutensis Kiritch. et Samyl.				+			+						
Cz. kanensis Kiritch. et Samyl								+					
Phoenicopsis irkutensis Dolud. et Rasskaz.				+	+		+						
Phoenicopsis angustifolia Heer			+										
Ph. samylinae Kiritch. et Moskv.			+										

Table 3. (Contd.)

1	2	3	4	5	6	7	8	9	10	11	12	13	14
Ph. markovitchiae Kiritch. et Shishkina						+							
Phoenicopsis sp.		+											
Leptostrobus laiflora Heer				+	sp.								
L. crassipes Heer	+												
Brachyphyllum sp.								+					
Pseudotorellia cf. *abaniensis* Kiritch.			+										
Schizolepidium gracile Heer	+												
Sorossacus sibiricus Heer	+												
Elatides ovalis Heer			+										
Samaropsis sp.	+		+										

The upper part of the Tyumen' Formation (its upper subformation) was recovered by many wells. In the Tomsk region, it is represented by a fine interbedding of siltstones, clays, and sandstones (Fig. 3), with plant remains discovered in the Tolparovo, Northern Chertalinka, Southern Koltogorovo, and Zapadnyi Luginets areas (Table 5).

In the Urengoi area, the Tyumen' Formation is overlain unconformably by the marine Abalakovskaya Formation. The lower part of this formation is represented by silty, brown-black claystones and grades upward into dark grey, fine calcareous claystones. The formation bears Callovian–Kimmeridgian ammonites (Mesezhnikov *et al.*, 1984; Braduchan, 1985), among which we found the late Callovian forms *Longaviceras* ex. gr. *filarum* Meled., *Camptonectes* (*Camptonectes*) ex. gr. *lens* (Sow.), and *Pleuromya* sp. (identifications by Shul'gina and Vavilov), which were yielded by the Abalakskaya Formation (Well 673, depth range 3678 - 3708 m).

In the sublatitudinal Ob' and northern Tomsk regions, the Abalakskaya Formation is replaced either by the coastal-marine Vasyugan Formation or by the terrestrial Naunak Formation. The Vasyugan Formation consists of claystones and dark grey, locally bituminous clays, sandstones, and siltstones. The formation contains Callovian–lower Kimmeridgian foraminifers, rare ammonites, and plant fossils in the terrestrial part of the section (Mesezhnikov *et al.*, 1984; Tat'yanin and Volkov, 1982). The Naunak Formation, which is the purely terrestrial analog of the Vasyugan Formation, is also a clayey unit with frequent inclusions of carbonaceous detritus and plant fossils (*Stratigraficheskii* ..., 1978). The latter were first studied by Bystritskaya (Bystritskaya and Tat'yanin, 1983; Bystritskaya *et al.*, 1992). We obtained the most complete paleofloral characteristics of the Vasyugan Formation and its analogs from wells of the sublatitudinal Ob' and Tomsk regions (Fig. 3). The species composition of all taphoflora groups from this part of the section differs greatly from that of the Tyumen' Formation (Table 6).

Thus, the specified taxonomic composition and stratigraphic levels of the western Siberian taphofloras could serve as a basis for the biostratigraphic subdivision and regional correlation of the Jurassic deposits, as well as for their dating using plant megafossils. In our opinion, this very difficult problem can be solved by revealing the evolutionary stages of the western Siberian paleoflora and by reconstructing its paleosuccessions. This will enable us to compare different western Siberian taphofloras with each other and with those from neighboring regions using their evolutionary levels rather than their taxonomic composition.

Analysis of the stratigraphic positions of the floral assemblages revealed a certain trend in their distribution throughout the Jurassic section and, thus, the general tendency of paleoflora evolution with time. There are three large stages in the evolution of the western Siberian paleoflora. Each is classified as a regional stratoflora (Kirichkova, 1982) that subdivides the Jurassic deposits into the Urengoi (Lower Jurassic), Tomsk (Middle Jurassic), and Naunak (Middle-Upper Jurassic) horizons, which bear their proper names. The Tomsk and Naunak horizons were subdivided into beds, and peculiar phytostratigraphic assemblages were used in the detailed regional correlation of different-facies deposits.

We recognized the Urengoi Horizon as unconformably resting on the Tampei Group and forming the basal part of the Jurassic succession. The floral assemblage of this group is similar to that of the Upper Triassic Trans-Ural Horizon (the Bulanash Beds) of the Eastern Urals, as is indicated by dominant horsetails of *Neocalamites* and *Paracalamites* genera (Kirichkova, 1990). The Urengoi Horizon unites the Beregovoi, the Yagel', and the main part (without the upper member) of the Kotukhta or Khudosei formations (Table 2). The key sections of this horizon are those of Urengoi Well 673

Table 4. Plant assemblage from the lower and middle subformations of the Tyumen' Formation

Species \ Locality, well, depth, m	Tomsk region					
	S.-Tambaevka, 76, depth 2848 - 2862	W.-Ostaninka, 447, depth 2748 - 2793	N.-Chertalinka, 400, depth 2998 - 3003	S.-Poselok, 1-R, depth 2956 - 3063	Verkhnezayachii, R-80, depth 2808 - 2873	Sublatitudinal Ob' region, Var'-Egan, 952, depth 3090 - 3110
Coniopteris simplex (L. et H.) Harris				+		
C. hymenophylloides (Brongn.) Sew.			+	+		
Coniopteris sp.		+		+		+
Cladophlebis suluctensis Brick						+
C. ex. gr. *whitbiensis* (Brongn.) Brongn.	+					
Raphaelia diamensis Sew.				+		+
Nilssonia baranovae Burak.			+			
Sphenobaiera aff. *uninervis* Samyl.					+	
Sphenobaiera sp.		+				
Leptotoma sp.		+				
Czekanowskia cf. *obiensis* Kiritch. et Samyl.				sp.	+	
Phoenicopsis angustifolia Heer		+				
Ph. varia Kiritch. et Trav.		+				+
Ph. samylinae Kiritch. et Moskv.	+				+	
Ph. cf. *taschkessiensis* Krasser						+
Phoenicopsis sp.				+		
Pseudotorellia sp.					+	

(depth range 4590 - 5060 m) and Prikoltogorovo Well 2-R (depth range 3150 - 3300 m) in the Tomsk region. The Urengoi Horizon is composed of coarse-grained sandstones (beds Yu_{9-13}) alternating with members of interbedded sandstones, clays, and carbonaceous-clayey rocks. The horizon is variable in its occurrence. It was recovered most completely in the Urengoi region. Southward, in the sublatitudinal Ob' and Tomsk regions, the horizon is either represented by its upper half (Table 2), which lies unconformably on Paleozoic deposits, or by weathering crusts. It disappears in the Gerasimovka and Zapadnyi Luginets areas of the Tomsk region (Fig. 3). The Urengoi stratoflora is characterized by horsetails of *Neocalamites* and *Equisetites* genera and by the rare ferns *Cladophlebis* sp., *Coniopteris* sp., *Todites princeps* (Presl.) Goth., and *Phlebopteris polypodioides* Brongn. Its gymnosperms are more diverse and include *Ginkgo ananievii* Bystr., *Sphenobaiera magnifolia* Aksarin, *Czekanowskia irkutensis* Samyl. et Kiritch., *Florinia* (?) sp., *Schizolepis* cf. *moelleri* Sew., and *Pityospermum* cf. *nanseni* Nath. The Lower Jurassic age of the Urengoi Horizon is evident from the absence of Middle Jurassic forms of the Siberian paleofloral province (Vakhrameev *et al.*, 1970) in the peculiar Urengoi stratoflora, from rare finds of Pliensbachian–Toarcian foraminifers (Table 2), and from the stratigraphic position of the horizon, which is similar to that of the marine Zima, Levinskii, and Dzhangoda formations of the Ust'-Enisei region. Moreover, the Urengoi stratoflora corresponds on an evolutionary level to the Bungarap stratoflora of the Kuznetsk basin (Kirichkova *et al.*, 1992), which we tentatively refer to the second half of the Early Jurassic.

The Tomsk Horizon, which either conformably overlies the Urengoi Horizon or lies unconformably on the Paleozoic deposits, comprises the upper (Radom) member of the Khudosei or Kotukhta formations and the Tyumen' Formation (Table 2). Bituminous claystones at the base of the horizon grade upward into alternating beds of grey sandstones, siltstones, and claystones with rare, thin interbeds of coals and conglomerates. The thickness of the horizon is about 500 m, but it reaches 800 m in the Urengoi region. The key sections of the horizon are those of Urengoi Well 673 (depth range 3740 - 4590 m), Southern Koltogorovo Well R-1 (depth range 2770 - 3120 m), and Tolparovo Well 1 (depth range 2670 - 3170 m). The horizon is wide-

spread. Its marine analogs in the northern region of western Siberia are an upper part of the Dzhangoda Formation and the Vymskoe, Leont'evskoe, and Malyshevka formations (*Stratigrafiya ...*, 1976).

The stratoflora characterizing the Tomsk Horizon differs greatly from the Urengoi stratoflora. It is still dominated by gymnosperms, but all groups have a more diverse taxonomic composition. This is especially true of Czekanowskiales, which are represented by 15 species of *Czekanowskia*, *Phoenicopsis*, and *Leptostrobus* genera (Tables 3 - 5) as well as Ginkgoals of *Ginkgo*, *Leptotoma*, and *Sphenobaiera* genera along with various fructifications of ancient Pinaceae: *Elatides*, *Sorosaccus*, *Samaropsis*, and others. Genus *Coniopteris*, which included the appearance of the Middle Jurassic species *C. maakiana* (Heer) Pryn., *C. nerifolia* Genk., *C. embensis* Pryn., *C. simplex* (*L.* et *H.*) Harris, and others, became very important among the ferns.

The sharp change in the taxonomic composition of the taphofloras from the lower part of the Tomsk Horizon, where *Coniopteris* (4 species) and diverse *Ginkgophytes* (up to 19 species) begin to play an important role, allows us to place the Lower-Middle Jurassic boundary at the base of the Tomsk Horizon, i.e., at the base of the Radom Member of the Khudosei or Kotukhta Formations, in accordance with the concept of the floral development across this boundary (Vakhrameev, 1969, 1982; Kirichkova *et al.*, 1992). Moreover, the confinement of taphofloras with diverse Czekanowskiales, *Coniopteris maakiana*, *Pseudotorellia abaniensis* Kiritch., and *Leptotoma sibirica* Kiritch. et Bat. (Table 3) to the lower part of the horizon allows us to classify this part of the section as *Coniopteris maakiana–Leptotoma*

Table 5. Plant assemblage from the upper subformation of the Tyumen' Formation

Species \ Locality, well, depth, m	Tomsk region				Sublatitudinal Ob' region, Les, R-285, 2630 - 2745	Urengoi region				
						673				
	Tolparovo, 1-R 2710 - 2724	W.-Luginets, 180, 2468 - 2485	N.-Chertalinka, 400, 2858 - 2866	S.-Koltogorovo, 1-R, 2800 - 2817		3741 - 3876	3950 - 4016	414, 3638 - 3681	411, 3858 - 3886	336, 3680 - 4600
Hepaticites sp.		+			+					
Neocalamites sp.						+				+
Equisetites asiaticus Pryn.	+				+				+	
E. beanii (Bunb.) Sew.				+		cf.				
E. lateralis (Phill.) Phill.		+				sp.			sp.	+
Phyllotheca sibirica Heer								+		
Coniopteris cf. *maakiana* (Heer) Pryn.	+	+			+					
C. cf. *depensis* E.Leb.		+					+	+	+	+
C. cf. *burejensis* (Zall.) Sew.					+					
C. hymenophyllodies (Brongn.) Sew.	+			+			+			
C. cf. *snigirevskiae* Tesl.		+								
C. simplex (L. et H.) Harris					+				+	
C. vialovae Tur.-Ket.			+							
C. cf. *spectabilis* Brick							sp.	sp.	sp.	+
Cladophlebis denticulata (Brongn.) Brongn.			+					sp.		
C. haiburnensis (L. et H.) Sew.				+						
Raphaelia stricta Vachr.				sp.		sp.	sp.	+		+
Czekanowskia jenissejensis Kiritch. et Samyl.		+	sp.		+					
Cz. cf. *obiensis* Kiritch. et Samyl.					+					
Phoenicopsis angustifolia Heer	+									
Phoenicopsis sp.		+		+						
Podozamites lanceolatus (L. et H.) Schimp.	+									
Schizolepis sp.		+						+		

Table 6. Plant assemblage from the Vasyugan Formation

Species \ Locality, well, depth, m	Tomsk region							Sublatitudinal Ob' region			
	Vesennyaya, 251	S.-Tambaevka, 76, 2592 - 2606	W.-Ostaninka 447, 2526 - 2560	Gerasimovka, 8, 2540 - 2588	W.-Luginets, 180, 2424 - 2438	Verkhnezayachii, 80, 2688 - 2727	S.-Koltogorovo, 1, 2703 - 2710	Var'-Egan, 952, 2865 - 2879	S.-Enitor, 165, 2638 - 2648	Koshil', 304, 2304 - 2340	Bachil, 127, 2645 - 2655
1	2	3	4	5	6	7	8	9	10	11	12
Lycopodites sp.				+							
Equisetites beanii (Bunb.) Sew.		sp.	+								sp.
E. lateralis (Phill.) Phill.				+					+		
Coniopteris latilobus Bystr.		sp.	+						sp.	sp.	
C. cf. *furssenkoi* Pryn.				+							
C. hymenophyllodies (Brongn.) Sew.	+						+				
C. vialovae Tur.-Ket.											+
C. simplex (L. et H.) Harris	+										+
C. aff. *maakiana* (Heer) Pryn.	+				+						
C. cf. *depensis* E. Leb.							+				
C. whitbiensis (Brongn.) Brongn.							+				
Raphaelia sp.										+	+
Nilssonia majskaja Bystr.	+		+	+							
N. urmanica Bystr.			+								
Nilssonia sp.			+								
Cycadolepis sp.								+			
Erethmophyllum sp.								+			
Leptotoma borealis Kiritch. et Trav. sp. nov.			+			sp.				sp.	sp.
Czekanowskia cf. *auliensis* Dolud.						sp.		sp.		+	
Cz. tomskiensis Kiritch. et Samyl.			+	+	+				+		
Cz. vera Kiritch. et Samyl.											+
Phoenicopsis atrichomus sp. nov.						+					
Ph. sibirica Kiritch. et Trav.		+	+			+					
Ph. aff. *taschkessiensis* Krasser		+									
Ph. cf. *dentata* Pryn.										+	
Podozamites eichwaldii Schimp.			+	+	sp.	sp.					
Brachyphyllum sp.											+

sibirica–Phoenicopsis irkutensis Beds. The thickness of the beds ranges from 25 to 140 m. They are traced throughout the Tomsk region and recovered from wells in the sublatitudinal Ob' and Urengoi regions (Table 2).

The poor paleofloral characteristics of the middle part of the Tomsk Horizon, corresponding to the lower and middle subformations of the Tyumen' Formation, are mainly a result of poor recovery. In this part, we defined in this part rare *Coniopteris simplex* (*L.* et *H.*) Harris, *Cladophlebis suluctensis* Brick, *Raphaelia*, *Nilssonia*, and much less diverse forms of *Czekanowskia* (one species) and *Phoenicopsis* genera (Table 4).

More diverse taphofloras of the upper part of the Tomsk Horizon were collected from a greater number of wells (Table 5). These taphofloras are from the upper subformation of the Tyumen' Formation and its analog, the Malyshevka Formation of the northern region of western Siberia. The taphoflora in this part of section is distinguished by predominant small-pinnuled ferns of

genus *Coniopteris* – *C. vialovae* Tur.-Ket., *C. simplex* (*L.* et *H.*) Harris, and *C. snigirevskiae* Tesl. – and by less diverse genera *Czekanowskia* and *Phoenicopsis* (Table 5). In general, the main components of these taphofloras are numerous *Equisetites Coniopteris* (7 species) and *Raphaelia* forms associated with *Czekanowskia jenissejensis* Samyl. et Kiritch. The restriction of these taphofloras by the upper part of the Tomsk Horizon and the persistence of their composition allow us to classify this part of the section as the *Coniopteris simplex–Raphaelia stricta* Beds. The thickness of these beds is 80 - 150 m, and in the northwestern areas, up to 400 m. The beds are easily traceable in many wells of western Siberia (Figs. 2 and 3); they are missing in the Gerasimovka and Fedyushkino areas of the Tomsk region (Figs. 2 and 3).

The Middle Jurassic age of the Tomsk Horizon is reliably constrained by specific Tomsk stratoflora corresponding to the Middle Jurassic floras of the Siberian paleofloral province (Vakhrameev, 1988). It is controlled by finds of Bajocian–Bathonian foraminifers in the Leont'evskoe and Malyshevka formations of the northern region of western Siberia (Bulynnikova and Yasovich, 1972) (Table 2) and by Bathonian foraminifers from the top of the Tyumen' Formation of the Urengoi and sublatitudinal Ob' regions (Tylkina and Komissarenko, 1977). In addition, the Tomsk stratoflora is comparable on an evolutionary level to the Itat stratoflora of the Kansk–Achinsk coal basin and to the Chernyi Etap stratoflora of the Kuznetsk coal basin (Kirichkova and Travina, 1990; Kirichkova *et al.*, 1992). The stratofloras are most similar at the level of the *Coniopteris maakiana–Leptotoma sibirica–Phoenicopsis irkutensis* Beds, corresponding to the Chernyi Etap taphofloras of the Kuznetsk basin and the Rybnaya taphofloras of the Kansk basin. We tentatively accept the Aalenian age for all of them. However, it should be noted that the Tomsk stratoflora evolved for a longer period, including the Bathonian, as compared to the Itat and Chernyi Etap stratofloras. At this stage, younger coenoses existed with small-pinnuled *Coniopteris*, *Raphaelia, Nilssonia*, and *Czekanowskiales* of different species composition.

The Naunak Horizon, unconformably overlying the Tomsk Horizon in some places (Table 2), corresponds in stratigraphic range to the Abalakovo Formation of the northern region of western Siberia, the Vasyugan Formation of the sublatitudinal Ob' region, and the Naunak Formation of the southern Tomsk region. The thickness of the horizon is up to 115 m. The key sections of the horizon are those recovered from wells in the Pervomaiskaya (Well 272, depth range 2525 - 2570 m), western Ostanino (Well 447, depth range 2510 - 2590 m), and Gerasimovka (Well 8, depth range 2490 - 2590 m) areas of the Tomsk region; the Bakhin area of the sublatitudinal Ob' region (Well 127, depth range 2645 - 2655 m); and the stratotypes of the Vasyugan and Abalakovo formations (*Stratigraficheskii* ..., 1978).

The Naunak stratoflora first reported by Bystritskaya (Bystritskaya and Tat'yanin, 1983) differs greatly from the Tomsk stratoflora in the species composition of all groups, although the relationship between the latter does not change (Table 6). Ferns are represented by different species of *Coniopteris*, consisting of *C. latilobus* Bystr., *C. vialovae* Tur.-Ket., *C. vsevolodii E.* Leb., and by new species of *Nilssonia–N. majskaya* Bystr. and *N. urmanica* Bystr. However, the most significant aspect is a renewal of species composition of genera *Czekanowskia* and *Phoenicopsis*, which are represented by forms unknown from the underlying deposits. In general, the taphofloras of the Naunak stratoflora reflect another, succeeding stage in the Siberian paleoflora evolution. The typical restriction of the Naunak taphofloras by the lower half of the horizon allows us to identify them as the *Coniopteris latilobus–Czekanowskia tomskiensis–Phoenicopsis sibirica* Beds. The thickness of the beds is 20 - 50 m.

The Callovian-Kimmeridgian age of the horizon as a whole is inferred from the foraminifer and ammonite occurrences in the sections of the Vasyugan and Abalakovo formations (Mesezhnikov *et al.*, 1984). The assignment of the *Coniopteris latilobus–Czekanowskia tomskiensis–Phoenicopsis sibirica* Beds to the Callovian–Oxfordian is based on the presence of foraminifers in the Vasyugan and Naunak Formations of the Tomsk region (Tat'yanin and Volkov, 1982; Bystritskaya and Tat'yanin, 1983). Presently, it is difficult to find a Siberian paleoflora analog for the Naunak stratoflora. It corresponds to the Chechuma stratoflora of West Yakutiya on an evolutionary level but differs sharply in its taxonomic composition and, therefore, is not comparable with it.

Thus, detailed analysis of the taxonomic composition of the plant fossils from the Jurassic deposits of western Siberia and their stratigraphic levels yielded the following results:

1. The revealed details in the composition of floral assemblages from the Jurassic of western Siberia have shown (Kirichkova, 1989) that they are dominated by gymnosperms, mainly by Czekanowskiales, conifers, and Ginkgoals. Ferns became more important at the end of Middle and beginning of Late Jurassic. The main components of the assemblage are *Czekanowskia, Phoenicopsis, Leptotoma, Ginkgo*, and *Sphenobaiera* genera. Ancient Pinaceae represented by fructifications and leaves are the dominant conifers. Ferns are represented by *Coniopteris, Raphaelia*, and, to a lesser extent, *Cladophlebis*.

2. The recognized stages in the Jurassic evolution of the western Siberian flora allow us to suggest the first phytostratigraphic scheme of the terrestrial and coastal-terrestrial Jurassic deposits; distinguish the regional Urengoi (Lower Jurassic), Tomsk (Middle Jurassic), and Naunak (upper Middle–lower Upper Jurassic) horizons; and define their ages and regional distribution. Peculiarities of the floral evolution during a partic-

ular stage served as the basis for the subdivision of the horizons into floral beds and the detailed, long-distance correlation of different-facies deposits in western Siberia.

3. Noticeable changes in the taxonomic composition of the Urengoi and Tomsk stratofloras, which are manifested in the species diversity of *Coniopteris* in taphofloras from the lower part of the Tomsk Horizon [*C. latifolia* Brick., *C. nerifolia* Genk., *C. maakiana* (Heer) Pryn., etc.], permit the Lower-Middle Jurassic boundary to be drawn at the base of the Tomsk Horizon, i.e., at the base of the Radom Member.

4. The regional correlation of the floral beds showed that parts of the succession, such as the Lower Jurassic in the Tomsk region, the upper half of the Middle Jurassic, and the lower part of the Vasyugan Formation, are incomplete. Jurassic sedimentation began and ended at different times in different western Siberian regions.

5. The Czekanowskiales are of particular stratigraphic significance for reliable interregional correlation within a single phytochoria. This indicates that only detailed study on the taxonomy of leading groups of the Siberian paleoflora is able to provide a solid basis for the development of a more refined phytostratigraphy and for wide chronostratigraphic correlations.

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Evolution of the biosphere

Chronological Relationship of Pliocene Deposits in Fluviatile Plains between the Prut and Southern Bug Rivers

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Abstract – Deposits of three Pliocene fluviatile plains between the Prut and South Bug rivers were studied in order to reconsider their disputed age determinations and chronological relations. Different ages of these deposits have been established, and their correlation with the Eastern Paratethys stratigraphic scale and magnetochronologic and stratigraphic scales of West European continental deposits have been established. The oldest deposits of the Stol'nicheny Formation are dated as the early Kimmerian (magnetic Chron 5). On a continental scale, they are close to the boundary between the Turolian and Ruscinian. The Kuchurgan Gravel is dated as the terminal early and middle Kimmerian (middle part of the Gilbert Chron) and also as the terminal early Ruscinian (MN zone 14). The Karboliya Beds correspond to the late Kimmerian (the end of Gilbert Chron) and to the late Ruscinian (MN zone 15).

Neogene continental deposits are widespread in the area between the Prut, Dniestr, and South Bug rivers. Their formation was initiated at the end of the middle Sarmatian. At that time, the sea regressed from the Galician Bay far to the south and left a wide depression covered with marine sandy and clayey deposits of the middle Sarmatian. As the marine basin regressed the original fluviatile plain gradually progressed to the south. In the middle Pontian, this plain covered the entire territory of Moldova and southwestern Ukraine. Its deposits were composed of cyclic gravel, sandy, and clayey sediments of fluviatile, lacustrine, and deltaic origin. This continental sequence is subdivided into two major parts: the Miocene (Balta Formation) and Pliocene. Deposits of older fluviatile plains, containing "Carpathian pebbles" (jaspers, silified argillites, and others) and alluvium from the higher terraces of the Prut, Dniester, and Danube rivers are referred to the Pliocene (Khubka, 1981; Bukatchuk *et al.*, 1983).

Deposits of Pliocene fluviatile plains are preserved on the most elevated watershed areas. Only in the south of the Prut region are they submerged below the sea level. There are three fields of old fluviatile deposit occurrences (Fig. 1). These are areas between the South Bug and Dniester rivers ("Kuchurgan Gravel"), between the Dniester and Prut rivers (Stol'nicheny sequence), and between the Prut, Danube, and Yalpug rivers, where deposits bear mammals of the "Moldavian Roussellion" (Bilinkis, 1987). The latter have been named differently by different authors as the lower Porat Alluvium (Konstantinova, 1967), the Karboliya Beds (Roshka and Khubka, 1964), etc. Here, we use the term "Karboliya Beds". Old fluviatile deposits have been studied to varying extents. The Stol'nicheny Alluvium is so far the least known. There are different opinions concerning the chronological relationships of the above three sequences as well as their correlation with the regional stages of the Eastern Paratethys and the continental zonation scale of P. Mein (1990).

Stratigraphic relations between the Kuchurgan Alluvium and the Karboliya Beds are interpreted in two different ways. First, the Kuchurgan deposits are assumed to be older than the Karboliya Beds (Schevtschenko, 1965; Konstantinova, 1967; Korotkevich, 1988; Topachevskii, and Nesin, 1989). According to the second viewpoint, these deposits are synchronous (Lungershausen, 1938; Chepalyga, 1967; Khubka, 1982; Gabuniya *et al.*, 1986; Bukatchuk, 1986; Bilinkis, 1987; and others).

The Stol'nicheny deposits are considered either older than the Kuchurgan deposits and Karboliya Beds (Gabuniya *et al.*, 1986; Khubka, 1987) or synchronous with them (Bilinkis, 1987).

The goal of the present study is to refine the stratigraphic position of the three sequences under consideration.

KUCHURGAN ALLUVIUM

We studied Kuchurgan deposits in the right bank of the Kuchurgan River valley in the natural outcrops and in the quarries near the villages of Novaya Andria-

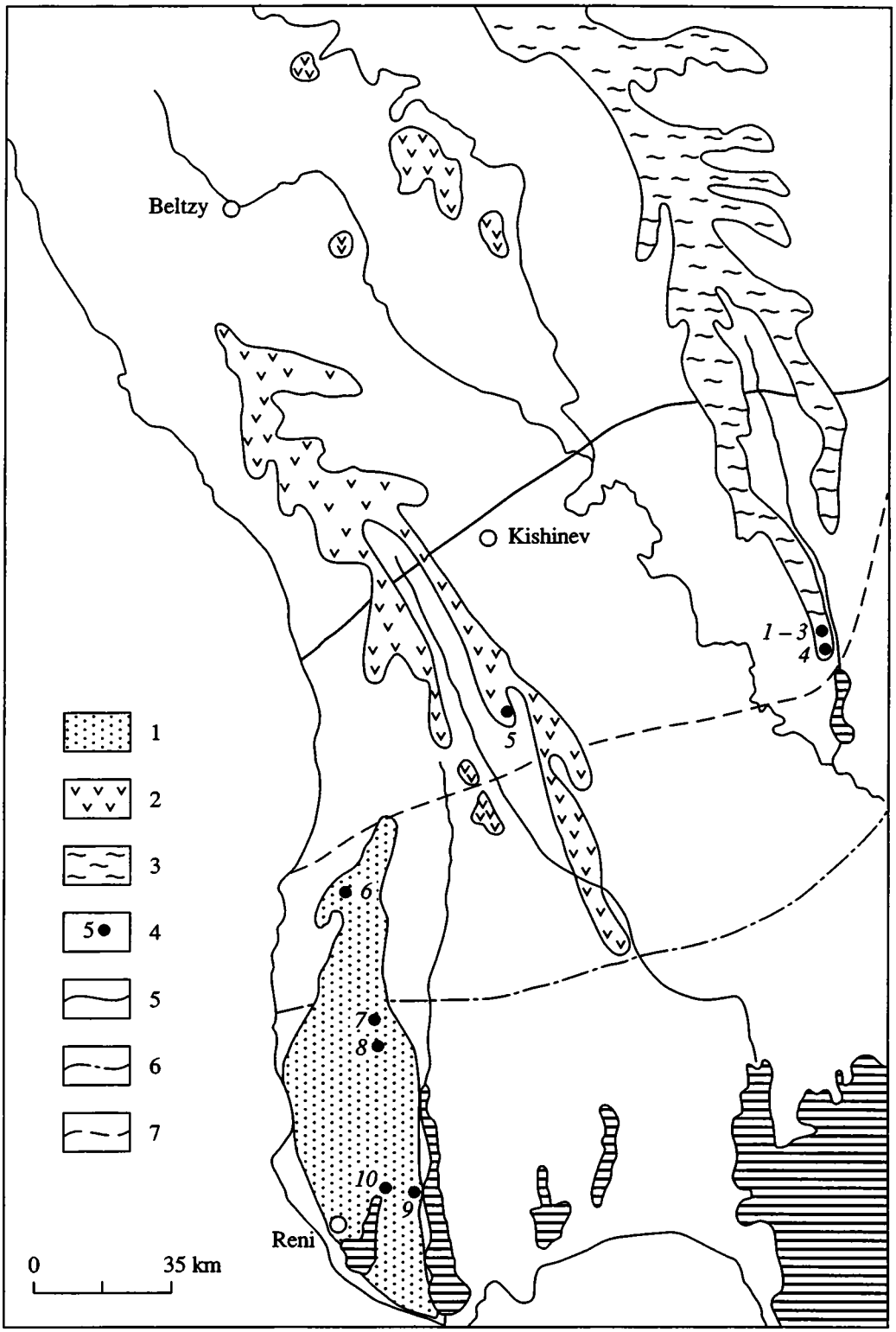

Fig. 1. Distribution of deposits of Pliocene fluviatile plains between the Prut and South Bug rivers.

1 – Karboliya Beds, 2 – Stol'nicheny Alluvium, 3 – "Kuchurgan Gravel", 4 – Localities of fossil mammals: *1* – Grebeniki 2, *2* – Nikol'skoe, *3* – Uyutnoe, *4* – Novaya Andriashevka, *5* – Mugureny, *6* – Lucheshty, *7* – Budei, *8* – Musait, *9* – Kotlovina, *10* – Novaya Etuliya.

Inferred coastlines for: 5 – Late Sarmatian Sea, 6 – Meotic Sea, 7 – Early Pontian Sea.

Distribution of fluviatile formations after Bilinkis (1987).

shevka, Uyutnoe (Konstantinovka), Slavyano-Serbskaya, Velikaya Ploskaya, Anastasievka, Novopetrovka (the stratotype of the Kuchurgan faunal assemblage), and Stoyanovo, and also in quarries near the villages of Savchenskoe, Voinichevo, and Ivanovka on the left bank of the Kuchurgan Valley.

In all the sites mentioned, the Kuchurgan deposits overlie various horizons of the Balta Formation. They have a thickness of 10 to 15 meters and are represented by medium- to fine-grained sands with common lenses of gravel and pebbles of Carpathian origin and rare clay lenses. The deposits often yield remains of larger and smaller mammals and rare shells of freshwater mollusks.

Small mammal remains were collected in quarries near the villages of Novaya Andriashevka and Grebeniki. They were also collected in smaller amounts in quarries near the villages of Nikol'skoe and Uyutnoe (Konstantinovka) (the table).

Grebeniki 2. The sand quarry of this locality is on the right bank of the Kuchurgan Valley, Grebeniki village (Odessa region), two km from the center of the village of Grebeniki near a cattle-breeding farm. The Kuchurgan Alluvium consists of medium-grained sands and gravels. They occur above the fine-grained clayey sands of the Balta Formation. The exposed thickness is up to 14 meters. The quarry bottom is almost at the top level of the Balta deposits. However, the contact between the Kuchurgan and Balta deposits has not been observed. Mammal remains were collected from a meter-thick lens of poorly sorted gravel, one or two meters above the quarry's bottom.

List of small mammals from the studied localities

List of forms	Mugureny	N. Andriashevka	Grebeniki 2	Nikol'skoe	Uyutnoe	Musait 7	Budei	List of forms	Mugureny	N. Andriashevka	Grebeniki 2	Nikol'skoe	Uyutnoe	Musait 7	Budei
RODENTIA								*Tamini indet.*		2	3				
Promimomys sp. 1	1							*Pliopetaurista* sp.	3	6	1				2
Promimomys sp. 2		1						*Blackia polonica* Mein		2					
Promimomys insuliferus (Kowalski)		45						*Miomimus* sp.		2					
								Glis minor Kowalski		1					
Promimomys moldavicus (Kormos)			20	2	1		31	*Eliomys* cf. *intermedius* Friant						1	2
Promimomys sp. n.						12		*Castor* ex gr. *fiber* L.					1		1
Promimomys ex gr. *occitanus* (Thaler)							2	*Trogontherium* cf. *minus* Newton	3	1	4	4			
Trilophomys sp.						1		LAGOMORPHA							
Epimeriones sp.			1					*Ochotona antiqua* (Argiropulo et Pidoplitshka)	1	60	65	5	2	1	9
Pseudomeriones sp.		1						*Ochotona* cf. *eximia* Chomenko		16	8	1	1		4
Kowalskia polonica Fahlbusch		6	1					*Pliolagomys* cf. *gigas* (Argiropulo et Pidoplitshka)						2	
Kowalskia cf. *magna* Fahlbusch		5						*Prolagus* cf. *michauxi* Lopez	1	3	8				
Kowalskia sp.	1							Leporidae indet.							1
"Cricetus" aff. *kormosi* Schaub		12	7					*Alilepus* sp.		1			1		
"Cricetus" sp. (large form)							7	*Trischizolagus dumitrescuae* Radulesku et Samson		4	12	4	1		2
Odessamys simionescui (Kormos)						1	27	*Hypolagus* sp.	2						
Prospalax sp.	4	8	3				1	INSECTIVORA							
Nannospalax macovei (Kormos)	5	25	13	1	3	4	47	Soricidae indet.	1		7	1			5
Occitanomys adroveri (Thaler)		10						Blarini indet.		2					
Occitanomys sp.							1	Soriculini indet.		2					
Apodemus cf. *dominans* Kretzoi		5	1			4	1	*Amblicoptus* sp.	1	2					
Apodemus sp. 1		2	2					*Talpa* sp.	4	2	2			2	
Apodemus sp. 2		8						Total	28	233	158	18	10	28	143
Muridae indet.	1														

Novaya Andriashevka. The old, weedy quarry on the right bank of the Kuchurgan Valley is situated between the villages of Frunze and Pervomaisk, 0.5 km south of the steep road from the plateau to the village of Novaya Andriashevka. Here, there is an exposure of fine- and medium-grained quartz sands (4 or 5 meters thick) with gravel lenses. Remains of small mammals were collected from the gravel lens in the section's middle part.

Nikol'skoe. The sand quarry in the eastern outskirts of the village of Nikol'skoe exposes the Kuchurgan Alluvium consisting of fine- to medium-grained quartz sands with rare gravel lenses. The exposed thickness is 5 to 6 meters. There is no visible contact with underlying deposits of the Balta Formation. In a ravine situated south of the quarry, the Balta sands rest on upper Sarmatian deposits with marine molluscan fauna. Mammal remains were collected from a gravel lens two meters above the bottom of the quarry.

Uyutnoe (Konstantinovka). The sand quarry of the village of Uyutnoe is situated two kilometers south of the Nikol'skoe Village quarry, between the villages of Uyutnoe and Konstantinovka on the plateau surface. There is an outcrop about 3 meters thick of medium- to coarse-grained sands with pebble and gravel lenses. Mammal remains were collected in the middle part of the exposure. The pelvic bone of a mastodon was also found here.

STOL'NICHENY ALLUVIUM

Comprehensive descriptions of these deposits have been presented by Khubka (1982) and Bukatchuk (1985).

We inspected fluviatile deposits of the Stol'nichehy Formation together with Moldavian geologists P.B. Bukatchuk, I.V. Blyuk, P.E. Guivan, and A.N. Lungu in a number of outcrops and quarries in the middle and central parts of the distribution area. The exposures are situated near the villages of Viveritsa, Leordoe, Pitushka, Gaureny, Kabaeshty, Selishte, Stol'nicheny, Zloti railway station, Novye Kirkaeshty, Sagaidak, Mugureny (Surik), and Troitskoe. The deposits are mainly composed of fine- to medium-grained quartz sands with lenses of clays, silts, gravel, and pebbles containing Carpathian material. The Stol'nicheny Alluvium occurs above various horizons of the Balta Formation. It overlies marine deposits of the lower Pontian only in the south. Its maximum thickness is to 40 - 50 m (Bukatchuk, 1985).

Bukatchuk (1985) distinguished several levels composed of the Stol'nicheny Alluvium that represent heterochronous (within the Kimmerian regional stage) terraces of the paleo-Prut and paleo-Dniester rivers. Bilinkis (Bukatchuk *et al.*, 1983) considers all these deposits as synchronous but "displaced to various hypsometric levels by Pleistocene neotectonic movements" (p. 39).

Mammal remains are extremely scarce in the Stol'nicheny Formation. This is a problem when trying to resolve different viewpoints concerning the age of its various parts that occur at different hypsometric levels.

We were able to collect some materials representing small mammal remains from the Mugureny sequence (the table) located in an old but sometimes used sand quarry on the right bank of the Skunosa River, which is the left tributary of the Kogil'nik River between the villages of Mugureny (Surik) and Sagaidak. The observed thickness of the deposits is 5 - 6 m. Mammal remains were collected in a gravel lens in the middle part of the section.

KARBOLIYA BEDS

The history of the research, occurrence conditions, lithology, and faunal characteristics (freshwater mollusks and mammals) of the Karboliya Beds have been thoroughly discussed (Khubka, 1982, 1987; Bilinkis *et al.*, 1987). Thus, we will be brief in our description of these deposits.

The Karboliya Beds rest on red weathering crust that developed after sandy and clayey deposits of the lower Pontian. In some places the weathering crust is eroded, and the Karboliya Beds rest directly upon the Pontian deposits. All the sequences dip southwest. For example, near the village of Tatareshty, the top of the Pontian has an absolute elevation of +158 - 160 m, but +140 m near the village of Lucheshty, +80 - 85 m in the vicinity of the village of Musait (Bilinkis, 1987), and −20 m near the village of Etuliya (Khubka *et al.*, 1983).

We observed the Karboliya Beds overlying the Pontian weathering crust near the villages of Dermenzhi (16th km of the Chumai-Tatareshty road), Budei (11th and 13th km), and Musait (5th to 6th km).

In the north, the Karboliya Beds are overlain by younger loesslike deposits or by red and brown buried soils of different age. In the south, they are overlain by younger deposits of fluviatile plains containing shells of sculptured Unionidae (the upper Poratian, according to Konstantinova and Poratian after Khubka) or by fluviatile deposits of Pliocene and Quaternary terraces of the Prut and Danube rivers. The maximum thickness of the Karboliya Beds is 60 m in the north and 110 m in the south (Valeny Village) (Khubka, 1982).

The Karboliya Beds represent constratous alluvium, sometimes including lacustrine facies. Up to eight alluvial cycles are recognized within them (Khubka, 1982, 1987 and others).

In sequences situated along the Bol'shaya Salcha Valley, the Karboliya Beds can be subdivided into two parts. The lower one predominantly includes horizontally stratified sandy and clayey deposits with rare layers of gravels mainly consisting of clayey pebbles and carbonate concretions. The upper part is substantially sandy and cross-bedded with a large number of layers and lenses of pebbles and gravels containing a considerable amount of Carpathian material. Some studies have used fluviatile cycles to correlate distant sections. This practice is incorrect because a number of cycles can vary even in different walls of the same ravine.

Freshwater mollusks are rare in these deposits but evenly distributed throughout the sequence. Shells of the following smooth and plicated Unionidae have been found here (Khubka, 1987): *Potomidae saratae* (Teiss.), *P. pannonica* (Neum.), *P. sibinensis* Pen., *P. stoliczkai* (Neum.), *P. sanderbergeri* (Neum.), *P. bogatscheevi* (Gr.-Beres.), *Eulimnium sturdzae* (Cob.), and *Plicatibathia flabellatiformis* (Mikh.) This fauna enabled Andreescu (1975), Khubka (1987), and other scientists to correlate the Karboliya Beds with deposits of the lower Romanian (the *Viviparus bifarcinatus* Zone NSM_{10}) occurring on the right bank of the Prut River and in the Dacian basin.

Mammal remains occur throughout the sequence, though the main finds are from the upper part.

We studied the Karboliya Beds along the valley of the Bol'shaya Salcha River. The deposits are exposed in a number of ravines along the right bank of the valley, between the villages of Vinogradovka (Vulkaneshty district) and Tatareshty (Kagul district) of Moldova. Mainly small mammal remains were collected near the villages of Budei and Musait.

Budei. The studied ravine is located on the right bank of the Bol'shaya Salcha Valley near the northern outskirts of the village of Budei. The road leading from Chumai to Tatareshty crosses it at the 13th km. Here, the lower Pontian deposits have a visible thickness of about 50 m. Above these deposits, there is a 3-m-thick red weathering crust. The lower sequence of the Karboliya Beds occurs above the weathering crust. It is represented by white and light yellow fine-grained quartz sands with rare gravel lenses. They have a visible thickness of 17 - 18 m. The upper part of the section is covered with talus. Small mammal remains were collected in gravel lenses located 1.5 to 5 m above the base of the Karboliya Beds.

In a similar geological setting fauna of small mammals was discovered by Alexandrova (Sadchikova *et al.*, 1986) in the southern outskirts of the village of Musait at the 5th km of the Chumai - Tataresty road (Musait 5).

Musait 7. The right bank of the Bol'shaya Salcha valley near the village of Musait is dissected by a complex ramifying system of ravines. In the center of the village at the 7th km of the road there is a junction of three major ravines.

The Karboliya Beds are represented here more completely than in sections near Budei and in the southern outskirts of the village of Musait (5th and 6th km of the road). The deposits rest here upon an eroded surface of the lower Pontian and are overlain by red and brown paleosoils. The thickest sequence of the Karboliya Beds is exposed to the right of the above-mentioned ravines. The lower member is 25 - 30 meters thick, and the upper one has a thickness of 15 - 17 meters. In places, they are divided by eroded red and brown buried soil. Remains of small mammals were collected in a gravel lens in the middle part of the upper member, 10 m below the top of the Karboliya Beds.

PALEONTOLOGICAL MATERIALS

A list of forms from the studied localities is given in the table.

In order to solve stratigraphic problems, most of our attention was paid to the study of voles. Most important in estimating the evolutionary advantages of Arvicoline species are the following features: differentiation of the anterior part of M_1 and posterior part of M_3, degree of development of the enamel islets, number of molar roots, and the degree of hypsodonty quantitatively correlated with the height of dentine tracts.

Collections from Musait 5 and Lucheshty VII of Alexandrova (1989), from Etuliya of Shushpanov (1985), and from Kotlovina (lower beds) of Topachevsky and Nesin (1989) were studied for comparison.[1] We also used the material collected in the lower Pontian deposits of the Vinogradovka locality, the Bolgrad district, and the Odessa region (collection GIN No. 1093).

Below, we present a description of voles from the localities studied.

Mugureny. The single first lower molar (M_1) of *Promimomys* sp. (Fig. 2, *1*) from this site is a rather worn tooth that had already lost an enamel islet and has extremely low dentine tracts.[2] The sum of three tracts (C), namely the outer tract of the anteroconid and the outer and inner tracts of the posterior lobe is equal to 0.23 mm. Thus, the hypsodonty degree of the form is more advanced than that of *Promimomys* sp. from the Vinogradovka locality (C = 0.1 mm).

Novaya Andriashevka. The *Promimomys insuliferus* (Kow.) (Fig. 2, *2*) from this locality is distinctly more evolved than the Mugureny form. The two forms are apparently separated by a considerable time interval. The sum of the tracts in the first lower molar ranges here from 0.52 to 0.7 mm (n = 5), with a mean value of 0.62 mm versus 0.23 mm measured in the case of the vole from Mugureny. This form has a somewhat more differentiated M_1 anterior part than the type forms from Podlesice in Poland and the southern Russia localities of Antipovka and Chugunovka (Agadjanian and Kowalski, 1978). This may indicate the slightly younger age of the former.

It is important to note that, apart from quite numerous remains of *P. insuliferus*, a single M_3 of a larger vole was encountered in the locality. The dimensions of the tooth lie beyond the variation range corresponding to teeth of *P. insuliferus*, and the dentine tracts are also very low.

Grebeniki 2. Two forms of voles were found here. The smaller one, *Promimomys* sp., is comparable in size to *P. insuliferus*, but unfortunately its remains are very

[1] We mean the bone-bearing bed from the seventh fluviatile cycle of the Lucheshty section, according to Alexandrova (1989).

[2] Height of dentine tracts was measured from the conventional line drawn at the crown base normal to the long axes of the synclinal folds through the lowermost point of the enamel in the posterior external synclinal fold.

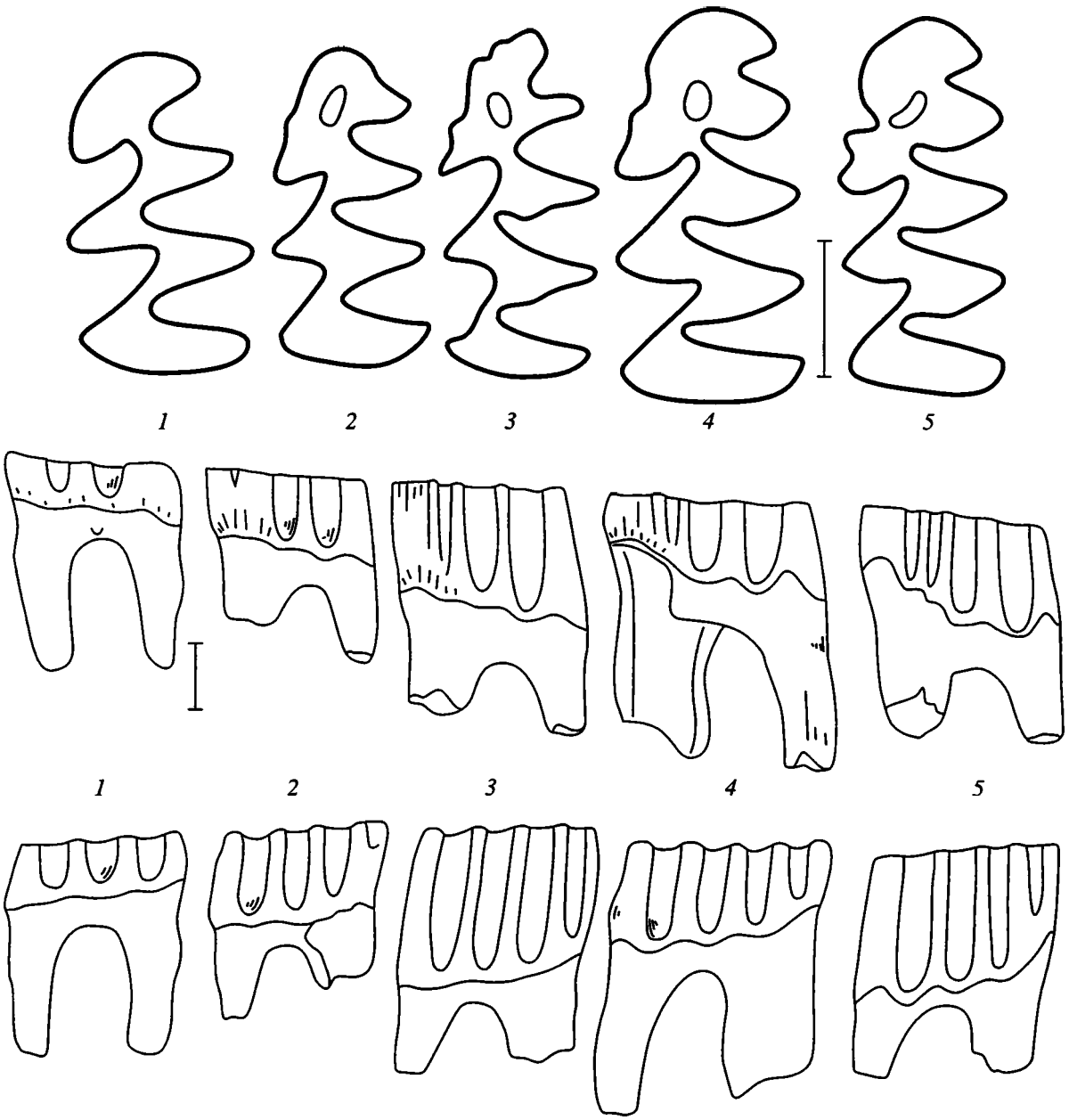

Fig. 2. Arvicolid molars of the *Promimomys* genus.

(*1*) *Promimomys* sp., Mugureny; (*2*) *P. insuliferus*, Novaya Andriashevka; (*3*) *P. moldavicus*, Grebeniki 2; (*4*) *P. moldavicus*, Budei; (*5*) *P.* sp. nov., Lucheshty VII. Upper row – occlusial surface, middle row – external view, lower row – internal view. Scale bars – 1 mm.

fragmentary. The form is slightly more hypsodont than the species from Novaya Andriashevka. The other, larger vole, *Promimomys moldavicus* (Kormos) (Fig. 2, *3*), already has a well differentiated M_1 anterior part; however, its tracts are still low (C = 0.7, 1.0, 1.2). This form is most similar to *P. moldavicus* from the Romanian locality Malusteny. Voles from Nikol'skoe and Uyutnoe represent the same stage of evolution.

Budei. Material from this locality contains remains of two Arvicolines. *Promimomys moldavicus* is smaller (Fig. 2, *4*). It is somewhat more hypsodont (C = 1.2, 1.3) than *P. moldavicus* from Grebeniki 2 and more similar to forms from the Ptolemais 3 locality in Greece (van de Weerd, 1979; Fejfar *et al.*, 1990) and the Calta locality in Turkey (Sen, 1977). A similar form was described by Alexandrova (1986) from the Musait 5 locality. Remains of the larger vole are fragmentary. They are arbitrarily attributed to *P.* ex gr. *occitanus*.

Musait 7. The locality yielded remains of *Promimomys* sp. nov. (Fig. 2, *5*). It is more hypsodont than

the Budei form and is identical to voles from the lower and middle fossiliferous beds of the Etuliya locality (Shushpanov, 1985), lower beds of Kotlovina (Topachevsky and Nesin, 1989), and Lucheshty VII (Alexandrova, 1989). The sum of the M_1 tracts varies from 1.5 to 2.1 mm. In its level of evolution, the form is very close to voles from the Weze locality in Poland (Sulimski, 1964), Czarnota 2 in Hungary (Kretzoi, 1962), and Orrios 3, 7 in Spain (Fejfar *et al.*, 1990).

COMMENTS ON SYSTEMATICS

In the last few years, it has been suggested that the boundary between the early and late Ruscinian be placed at the replacement level of *Promimomys moldavicus* (Scm.) by *Mimomys davakosi* van de Weerd (Fejfar and Heinrich, 1990). It was believed that these forms constitute a phyletic line. Radulescu and Samson (Radulescu *et al.*, 1989) consider them to be identical, with *Mimomys davakosi* being a synonym of *Promimomys moldavicus*.

In our opinion, the type materials on both species as well as the *Mimomys occitanus Thaler* are members of the same *Promimomys* genus of brachiodont rooted voles with an anterior enamel islet in M_3. Moreover, the discrimination of *P. moldavicus* from *P. davakosi* is questionable and requires specific evidence. At the moment, it is impossible to compare both forms in detail. In spite of adequate revisions (Radulescu *et al.*, 1989; Fejfar *et al.*, 1990), different measuring systems and inevitable size distortions in published illustrations make it difficult to compare dentine tract heights. Nevertheless, the structure of the M_1 anteroconid confirms that "*Mimomys*" *davakosi* is more advanced than *Promimomys moldavicus*. At the same time, the heights of the dentine tracts in M_1 of *M. davakosi* (van de Weerd, 1979; Fejfar *et al.*, 1990) and *P. moldavicus* (Radulescu and Samson, 1989) range from 0.8 to 1.2 mm. These values are close to those of *P. moldavicus* from Grebeniki 2. Thus, we are faced with the difficulty of discriminating *P. moldavicus* from *P. davakosi*. It is possible that they are conspecific. The most recent species in the *P. moldavicus* phylogeny is the form from the Musait 7, Luchesty VII, Etuliya (lower and middle beds), and Kotlovina (lower beds) localities. This form may warrant a separate specific name.

Summarizing the available data on voles from the considered sites, we may conclude that they represent sequential evolutionary stages as follows: primitive *Promimomys* sp. from Mugureny; *P. insuliferus* from Novaya Andriashevka; *P. moldavicus* from Grebeniki 2; slightly more advanced *P. moldavicus* from Musait 5 and Budei; *P.* sp. nov. from Luchesty VII, Etuliya (lower and middle beds), Kotlovina (lower bed), and Musait 5. A vole from the lower Pontian site Vinogradovka precedes all these forms.

Noteworthy is the presence of a second *Promimomys* form along with *P. insuliferus* already at the level of Novaya Andriashevka. Likewise, *P.* ex gr. *occitanus* occurs together with *P. moldavicus* in the Budei locality. This indicates the existence of several Arvicoline lines starting at a rather early stage in the group's history.

FAUNAL ASSEMBLAGES OF SMALL MAMMALS

Faunal groups accompanying various *Promimomys* forms undergo certain changes through time, in both quantitative ratios of different groups and the appearance or disappearance of certain species and even genera.

The quantitative relations between small mammals from the Mugureny locality is difficult to evaluate because of limited material. Most remains belong to mole rats, while other forms are almost equally represented. Noteworthy is the presence of the *Kowalskia* genus among them.

The diverse fauna of Novaya Andriashevka includes 26 species of 22 genera and 12 families. As noted, according to the evolutionary state of *Promimomys*, this fauna is separated from the Mugureny assemblage by a long time interval. At this stratigraphic level, the dominating *Ochotona* genus (34%) deserves attention. The second in abundance is the *Promimomys* genus (19%), and the amount of mole rats is significant (14%). Among hamsters, there are *Pseudomeriones* forms and two species of *Kowalskia*. Fauna of Muridae (10%) are represented by at least four species, including *Occitanomys adroveri*.

Faunas of Grebeniki 2, Nikol'skoe, and Uyutnoe with *Promimomys moldavicus* are similar to the fauna of Novaya Andriashevka in taxonomic composition but somewhat less diverse. The presence of *Epimeriones* is noted. These faunas are also similar to each other in quantitative proportions of species, though, in Grebeniki 2, Nikol'skoe, and Uyutnoe, the amount of pikas is higher (51%), but the abundance and diversity of Muridae is lower (6% and only two species, respectively).

Faunas of Budei and Musait 5 substantially differ from preceding assemblages in quantative proportions of forms. Pikas cease to be dominant (9%), being replaced by mole rats *Nannospalax macovei* (33%) and voles *Promimomys* (23%). There are also some differences in taxonomic composition: remains of *Kowalskia* forms are absent, and along with remains of large "*Cricetus*," bones of the small hamster *Odessamys simionescui* are met.

The youngest fauna of the Karboliya Beds with a very advanced species of *Promimomys* is known in the Musait 7, Lucheshty Vii, Etuliya (lower and middle beds), and Kotlovina (lower beds) localities. In addition to dominant forms of *Promimomys*, it is characterized by the appearance of the *Pliomys* and *Dolomys* genera. The assemblage from Musait 7 includes *Pliolagomys* and *Trilophomys* forms, encountered for the first time.

Three stages in fauna development during the time interval under consideration can be recognized according to the quantitative proportions and taxonomic composition of small mammals. The earliest one is tentatively recognized only in the fauna of the Mugureny locality. It corresponds to the time of the formation of the formation of the Stol'nicheny Alluvium. The stage is distinguished by almost equally represented groups except Spalacidae, but additional data is necessary to define the time limits of this stage. The second stage coincides with the formation time of the Kuchurgan Alluvium (Novaya Andriashevka, Grebeniki 2, Nokol'skoe, and Uyutnoe localities). A significant role of *Ochotona* forms, which became absolutely dominant at the end of the stage, is typical of the second stage.

In the Budei, Musait 5 and 7, Lucheshty VII, Etuliya, and Kotlovina localities, the Karboliya Beds were formed during the third stage. The stage began with the inversion of the Lagomorpha to Arvicolidae ratio forms. At the end of this time, interval the latter group became completely predominant. The lower boundary of the third stage is also marked by an extinction of the *Kowalskia*, *Prolagus*, and *Epimeriones* genera. In the middle of the stage, there is another distinct event: the appearance of the new genera *Pliomys* and *Dolomys* of the Arvicoline.

POSITION OF FLUVIATILE SEQUENCES IN THE MAGNETOCHRONOLOGIC SCALE

There is no serious disagreement regarding the position of the Karboliya Alluvium in the magnetochronologic scale. It is correlated with the upper part of the Gilbert Chron. Differences lie in the fact that some scientists place the lower boundary at the top of the Kochiti Subchron (Alexandrova, 1989), and others place it inside it (Khubka *et al.*, 1983). The upper boundary is considered to either coincide with that between the Gilbert and Gauss chrons (Alexandrova, 1989) or to coincide with the lowermost part of the Gauss Chron (Sadchikova *et al.*, 1983; Khubka *et al.*, 1983).

Paleomagnetic data is available for five sequences of the Karboliya Alluvium. In three of them (Tatareshty, Baimakliya, and Novaya Etuliya), a zone of normal polarity is registered in the lower part of the Karboliya Beds. It is correlated with the second half of the Kochiti Subchron. Accordingly, the lower boundary of the Karboliya Beds was placed inside the subchron (Khubka *et al.*, 1983). The reversely magnetized fossiliferous beds in the Musait 5 sequence (Sadchikova *et al.*, 1986) does not contradict this dating. In the uppermost parts of the Lucheshty (Sadchikova *et al.*, 1986) and Tatareshty sequences (Khubka *et al.*, 1983), a zone of normal polarity was distinguished. It corresponds to the beginning of the Gauss Chron. These normally magnetized deposits, however, lack paleontological characteristics. Whether they belong to the Karboliya Beds or to the overlying upper Poratian (Konstantinova, 1967) or Poratian deposits (Khubka, 1987) is a question that needs to be studied more thoroughly.

Hence, it follows that the Karboliya fluviatile deposits may occupy an interval of the paleomagnetic scale from the middle part of the Kochiti Subchron to the end of the Gilbert Chron (Fig. 3).

There is no paleomagnetic data for the Stol'nicheny and Kuchurgan fluviatile deposits. The position of these sequences in the paleomagnetic scale can be indirectly inferred by comparison with evolutionary stages of *Promimomys* forms in sections correlated with the magnetochronologic scale.

The vole from Grebeniki 2, i.e., from the upper part of Kuchurgan deposits, is very close to the type form *Promimomys moldavicus* described from the Romanian Malusteny locality. The latter is dated as the earliest Romanian and corresponds to the beginning of the Kochiti Subchron.

The position of the Novaya Andriashevka and Mugureny sections within the magnetochronological scale was determined by interpolating sum values of Arvicoline dentine tracts between corresponding values for voles from Grebeniki 2 and Vinogradovka (the latter site is dated as the beginning of magnetic Chron 6). Taking into consideration the exponential increase of the hypsodonty of *Promimomys* molars from the older to the younger forms, we estimated that the Novaya Andriashevka locality, that is, the lower boundary of the Kuchurgan Alluvium, falls in the middle part of the Gilbert Chron between the Nunivak and Sidufial subchrons. The Mugureny site (Stol'nicheny Alluvium) is placed within the magnetic polarity Chron 5.

POSITION OF FLUVIATILE SEQUENCES IN THE STRATIGRAPHIC SCALE OF THE EASTERN PARATETHYS

An analysis of published papers shows many different ideas on how to correlate the considered fluviatile sequences with regional stages of the Eastern Paratethys. The Stol'nicheny Alluvium is placed at various stratigraphic levels ranging from the middle Pontian to the late Kimmerian, inclusive; the Kuchurgan Alluvium is placed from the early Kimmerian to the early Akchagylian; and the Karboliya Alluvium is placed from the early Kimmerian to the middle Akchagylian.

Using inferred magnetochronological intervals for the deposits in question and the fact that the Kimmerian ranges from the end of Chron 6 to the end of the Gilbert Chron (Semenenko and Pevzner, 1979), we may conclude that the Stol'nicheny Alluvium was deposited in the early Kimmerian, while the Kuchurgan Alluvium and Karboliya Beds were formed in the middle and late Kimmerian, respectively (Fig. 3).

We emphasize that our inference on the age of the Stol'nicheny and Kuchurgan alluviums cannot be applied to the entire area in which these deposits occur.

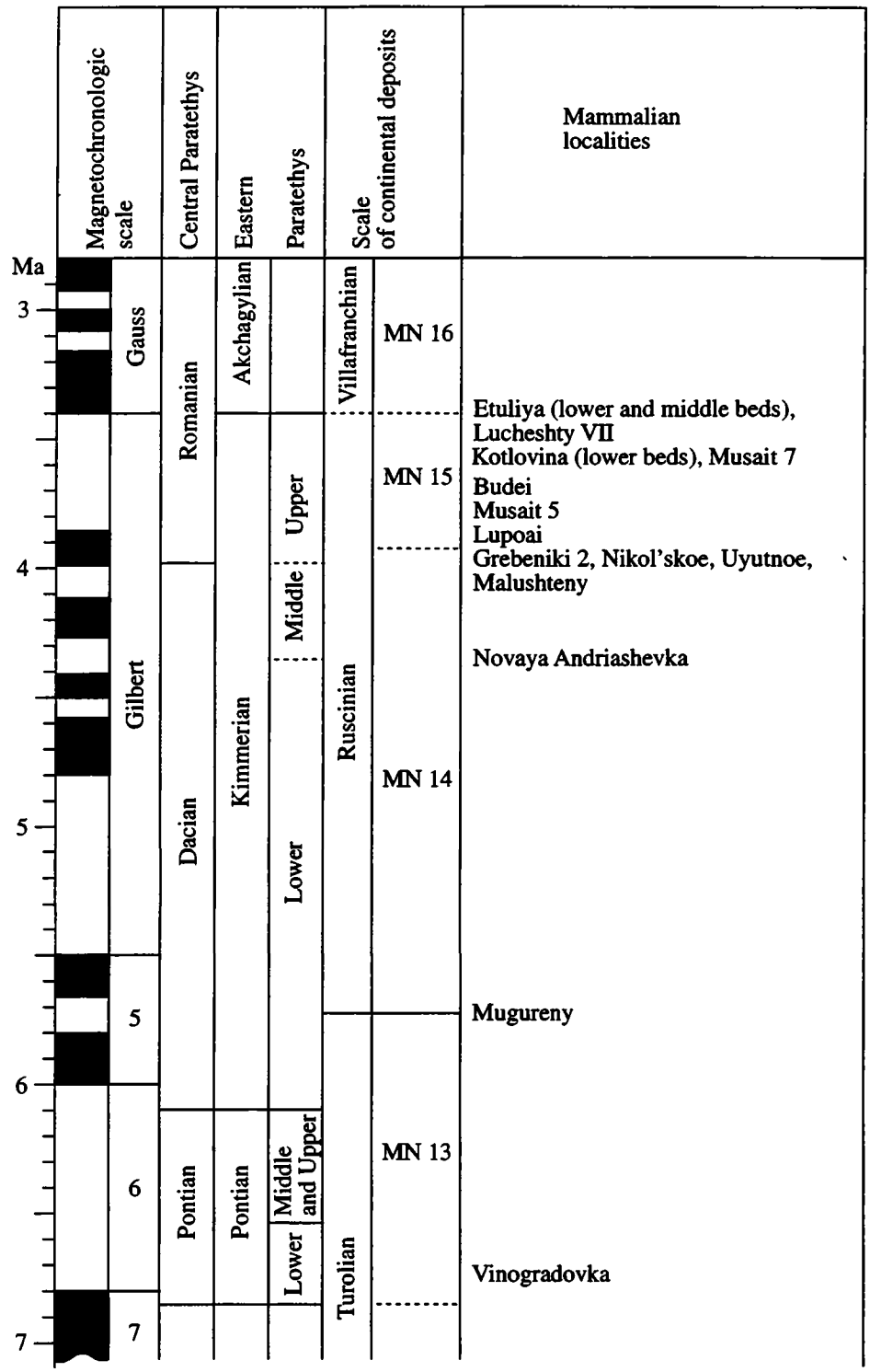

Fig. 3. Position of fossil mammal localities versus magnetochronologic and stratigraphic scales.

It is not improbable that further to the north they include some units considerably older than the Kimmerian.

As already noted, one of the main criteria of the subdivision of the supra-Sarmatian continental deposits into the Miocene and Pliocene parts was thought to be the appearance of "Carpathian" pebbles in the sequences. The time of this event in continental deposits of Moldova and South Ukraine has already been debated for several decades. Some Moldavian geologists suggested that the event occurred during the post-Pontian age (Bukatchuk

and Burdenko, 1969). There is some evidence, however, that the "Carpathian" pebbles had appeared in Moldavian territory already in pre-Kimmerian time. For example, in Stol'nicheny deposits with "Carpathian" pebbles near the village of Lozovo, a mandible of *Choerolophodon* sp. was found (Gabuniya *et al.*, 1986). This form of mastodon survived only until the end of the Meotian (Korotkevich, 1988). Therefore, the fossiliferous deposits are certainly pre-Pontian.

POSITION OF MAMMALIAN LOCALITIES IN THE CONTINENTAL DEPOSITS SCALE

The evolutionary level of the Arvicoline form from the Mugureny locality (Stol'nicheny Alluvium) strongly suggests that the fauna is not younger than the first half of the early Ruscinian, because the vole is more primitive than *Promimomys insuliferus*. The latter is typical for the second half of zone MN 14 (Mein, 1990). The lower time limit of this fauna is problematic. But if correlation of the Mugureny locality with the middle part of the magnetic polarity Chron 5 is correct, then the fauna is close in age to the boundary between the Turolian and Ruscinian (Pevzner and Vangengeim, 1986).

The co-occurence of *Promimomys insuliferus* and *P. moldavicus* similar to the type form in the Kuchurgan Gravel allows us to date it as the second half of the early Ruscinian (zone MN 14).

The boundary between zones MN 14 and MN 15 in Western Europe corresponds to the replacement level of *Promimomys moldavicus* by "*Mimomys*" *davakosi* (Fejfar and Heinrich, 1990). As was noted, it is extremely difficult to morphologically discriminate between these forms. That is why we cannot use paleontological criterion in the correlation of late Ruscinian faunas of Eastern and Western Europe and must use other methods, in particular, the magnetostratigraphic method.

The Karboliya Beds occupy a magnetostratigraphic position similar to that of the upper Alfambrian of Spain, i.e., to zone MN 15, between the Kochiti Subchron and the Gilbert – Gauss chrons boundary (Moissenet *et al*, 1990). Thus, the fauna from the Karboliya Beds corresponds to zone MN 15 in its chronological range, as defined by Moissenet *et al.* The Musait 5 and Budei localities can be attributed to the basal part of zone MN 15, and the Musait 7, Lucheshty VII, Etuliya (lower and middle beds), and Kotlovina (lower beds) localities can be attributed to the upper half of this zone (the Csarnotian of M. Kretzoi zonation as indicated by the appearance of the genera *Pliomys* and *Dolomys*).

CONCLUSION

The conducted study revealed different ages for deposits of three fluviatile plains occurring between the South Bug and Dniester and the Dniester and Prut rivers. Deposits of the Stol'nicheny Alluvium were formed during the early Kimmerian in the studied territory. On the continental scale, they correspond to the boundary interval between the Turolian and Ruscinian. On the magnetochronological scale, the Stol'nicheny deposits correlate to the middle part of magnetic Chron 5. The Kuchurgan Gravel was probably deposited at the very end of the early and in the middle Kimmerian, and is equivalent in age to the second half of the early Ruscinian (MN 14) of the continental scale. On the magnetochronologic scale, these deposits correspond to the middle part of the Gilbert Chron. The Karboliya Alluvium was formed in the late Kimmerian and correlates to the late Ruscinian (MN 15) on the continental scale, and with the time interval from the Kochiti Subchron to the upper boundary of the Gilbert Chron.

It should be noted that further studies are required to discover the paleontological criteria that could indicate the lower and upper boundaries of the early Ruscinian, as the available data is insufficient to correlate East and West European faunas.

Reviewer M.N. Alekseev

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New Data on Stratigraphy of the Upper Pliocene and Eopleistocene of the Lower Volga and Northern Caspian Sea Region

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Abstract – A detailed regional statigraphic scheme of the Pliocene and Eopleistocene is suggested. It is substantiated by characteristic faunal and floral assemblages and by paleomagnetic data.

For several years, we studied complex characteristics (biostratigraphic, lithologic-mineralogical, paleomagnetic, and others) of reference sections of the Akchagylian and Apsheronian deposits in the Lower Volga and northern Caspian Sea region. Mollusks, ostracodes, foraminifers, and spore and pollen assemblages, along with lithology, mineralogical composition, and paleomagnetic properties were studied in detail. These investigations resulted in a regional stratigraphic scheme (Zhidovinov *et al.*, 1982).

Further study of the above sediments and generalization of both new and previous data allowed us to revise details of the Pliocene and Eopleistocene stratigraphic scheme. We were also able to distinguish several independent biostratigraphic units of the beds' rank, which were classified as the lower and upper Pallasovka, the lower and upper Urda, the lower and upper Uzen' Beds of the Akchagylian deposits, and the lower and upper Tsubuk Beds of the Apsheronian sequence. Correlation of the stratigraphic subdivisions of different schemes is given in the table.

The lower Pallasovka Beds were found in many drilled sections (wells 25, 39, 40, 53, 100, 106, and others) of the Lower Volga and Northern Caspian Sea region (Figs. 1 and 2). Their stratotype is represented by the interval of 521 - 537 m of Well 40, drilled near the Khar'kovka settlement of the Pallasovka region of the Volgograd province (Zhidovinov *et al.*, 1987, 1990). The beds overlie sediments of different ages, from the middle Pliocene to the Early Cretaceous.

The lower Pallasovka Beds are distinguished as a unit according to lithologic and faunal data. They are represented mainly by clays with silty and sandy interlayers. Their freshwater molluscan assemblages include shells of *Pisidium, Unio, Valvata, Planorbis*, and other genera. In addition, the mass appearance of Akchagylian brackish-water *Clessiniola* forms (*C. utvensis, C. polejaevi, C. intermedia*, and others), together with single shells of the marine euryhaline species *Aktschagylia subcaspia* (Andrus.), was noted in the lower Pallasovka Beds. Freshwater species (representing *Cypria, Candona, Cytherissa*, and other genera) predominate among the microfauna forms, with their abundance decreasing upwards. Rare marine euryhaline foraminifers appear in these beds for the first time.

Within the central parts of the Caspian Sea region (Urda, Aralsor, Novo-Kazanka, and other areas), the lower Pallasovka Beds contain shells of brackish-water Clessiniola and rare marine euryhaline Mactrida forms. Freshwater gastropods of *Planorbis, Valvata, Bythina*, and other genera predominate in the Trans-Volga region near Saratov (Mokrous, Ershov, and other areas) and in the western part of the Caspian Sea region (Leninsk, Ilovatka-Nikolaevskii, and other areas). They occur together with bivalve mollusks *Dreissena polymorpha* Pall. and ostracodes of the *Cypria* and *Ilyocypris* genera. The character of the faunal assemblages changes upwards in the sequence. In the lowermost part of the lower Pallasovka Beds, mollusks are mainly represented by brackish-water *Dreissena* and *Clessiniola* genera and by abundant freshwater forms. Single, small mactridas *A. rubcaspia* (Andrus.) appear only in the upper part of the sequence.

The upper Pallasovka Beds were recognized in the same drill sections as the lower Pallasovka Beds (wells 25, 39, 40, 53, 100, and others). The interval of 492 - 526 m in the Well 40 section is described as their stratotype (Zhidovinov *et al.*, 1990). The beds are represented by clays with silty interlayers. Their thickness varies from 12 - 15 to 25 - 30 m. The beds were distinguished by the predominance of brackish-water faunal elements and the appearance of the marine euryhaline mollusks *Cerastoderma dombra* (Andrus.), *Aktschagylia ossoskovi* (Andrus.), and *Kirghizella inostranzevi* (Andrus.), as well as ostracodes of the *Cyprideis, Cypria, Ilyocypris*, and other genera. Within the Urda (Well 53), Aralsor (wells 100, 106), and other areas of the central part of the Caspian Sea region, molluscan assemblages

Correlation of the Upper Pliocene Stratigraphy of the Lower Volga and Northern Caspian Sea region

Stratigraphic units		Zhidovinov *et al.*, 1982;	Zhidovinov *et al.*, 1990
Regional stage	Substage	Beds	
Apsheronian	Upper	Zam'yany	Zam'yany
	Middle	Seroglazovka	Seroglazovka
		Tsubuk	Upper Tsubuk
			Lower Tsubuk
	Lower	Novo-Kazanka	Novo-Kazanka
Akchagylian	Upper	Aralsor	Aralsor
	Middle	Uzen'	Upper Uzen'
			Lower Uzen'
		Urda	Upper Urda
			Lower Urda
	Lower	Eruslan	Eruslan
		Pallasovka	Upper Pallasovka
			Lower Pallasovka

consist of marine euryhaline and brackish-water species of the *Cerastoderma, Aktschagylia,* and *Classiniola* genera. The taxonomic composition of mollusks in the upper Pallasovka Beds of the Trans-Volga region near Saratov is impoverished, and freshwater gastropods of the *Planorbis* and *Bithynia* genera appear among them. In the southeastern part of the Caspian Sea region (Saurbai area, sites 402, 388, and others), the upper Pallasovka assemblage includes the abundant bivalves *Aktschagylia subcaspia* (Andrus.), *A. karabugasica* (Andrus.), and *Cerastoderma dombra* (Andrus.), as well as the gastropods *Clessiniola utvensis* (Andrus.) and others.

The lower and upper Pallasovka Beds are characterized by a pollen assemblage of the steppe type, in which mainly herbaceous pollen (66.2%) of the Chenopodiaceae family predominates. Pollen of the meadow-steppe herbs is diverse. *Pinus* (23%) pollen predominates among the arborescent group; pollen of *Picea, Abies, Tsuga,* and broad-leaved *Carpinus, Quercus,* and *Acer* is also present. Spore forms include ferns and sphagnum mosses. The palynologic assemblage indicates temperate to cool climatic conditions. The temperature determined by Grichuk's (1969) method varied from –2 to –8°C in January, and from +15 to +21°C in July. The Palaeoarctic cryophilic ostracode species of *Cytherissa, Candona,* and *Ilyocypris* genera reveal climatic cooling.

The Eruslan Beds complete the lower Akchagylian sequence. The interval of 404 - 492 m in Well 40 near the Khar'kovka Settlement is defined as their stratotype. The beds are represented by clays with silty, sometimes sandy interlayers. They contain diverse fauna of brackish-water and marine euryhaline mollusks, ostracodes, and foraminifers. The most abundant species are *Cerastoderma vogdti* (Andrus.) and *C. pseudoedule* (Andrus.), among bivalves; *Potamides caspius* Andrus., among gastropods; *Polimnocythere tenuireticulata* (Suz.), *P. alveolata* (Suz.), and *Loxoconcha eichwaldi* Liv., among ostracodes; and *Elphidium incertum* (Chutz.) and *Ammonia beccarii* (L.), among foraminifers. Freshwater and brackish-water faunal elements predominate in the coastal and shallow marine facies, while marine euryhaline forms are very rare there. The first group is usually represented by bivalve species of *Dreissena, Unio,* and, rarely, *Rugunio* genera; gastropods of *Valvata* genus; shallow water ostracodes of *Cyprideis* genus; and foraminifers of *Ammonia* genus. Species of *Cerastoderma* and *Aktschagylia* genera are rare.

These beds contain a spore and pollen assemblage of the forest type, with a prevalence of arborescent forms (79.6%). The *Picea* and *Pinus* pollens predominate, and pollen of *Abies* and *Tsuga* is also abundant. Small-leaved forms (*Betula, Alnus, Salix*) predominate among hardwood species. Specimens of *Corylus* are most abundant among the broad-leaved forms. Herbaceous pollen is rare. The moss spore content (6.2%) increases, while the amount of fern spores decreases. The species composition of palynologic assemblage indicates a temperate–cool climate, with January temperatures ranging from +1 to –16°C, and July temperatures, from +16 to +20°C. Only the upper part of the beds is characterized by paleomagnetic data. Sediments here reveal the normal polarity that corresponds to the fragment of the Gauss p-orthozone (Zhidovinov *et al.*, 1989). The subordinate interval with reversed polarity in this part of the sequence may be correlated with the Kaena episode (2.8 - 2.9 Ma). The beds' thickness varies from 30 - 36 to 60 - 70 m, being less than 10 - 15 m above the domes.

The lower Urda Beds of middle Akchagylian age occur in all localities of the Lower Volga and the Caspian Sea region. Their stratotype is described as between 376 and 480 m of the Well 55 sequence, drilled near the Urda settlement in the Ural province (Figs. 1 and 2). The beds are represented by montmorillonite clays with silty interlayers in central parts of the province and by hydromica clays with sandy and silty interlayers in marginal areas. Their thickness varies from 14 to 80 m.

The beds are characterized by a rich assemblage of euryhaline and brackish-water, moderately thermophilic mollusks, ostracodes, and foraminifers. The assemblages of many bivalve and gastropod taxa include abundant marine euryhaline species of *Cerastoderma, Aktschagylia, Avimactra, Kirghizella, Potamides*, and other genera. New taxa of different rank that appeared here include genera *Avicardium, Caspimactra, Andrusovicardium*, and *Avardaria*, along with species *Aktschagylia striata* (Fedk.), *A. gedroitzi volkovae* (Fedk.), *Kirghizella pisum* (Andrtr.), and other forms. The lower Urda assemblage of abundant and diverse mollusks was found nearly everywhere except several peripheral zones, where freshwater species appear among them (the Kosheli Well 3 and others). According to Karmishina (1982), the lower part of the Urda Beds is characterized by the most diverse assemblage of marine and brackish-water euryhaline ostracodes and foraminifers.

The beds contain a pollen assemblage of the forest type, in which arborescent pollen content reaches 78.8% of the total amount, herbaceous pollen content is up to 22%, and spore content is not more than 10%. The coniferous pollen is dominant, being represented by specimens of *Pinus* (33.7%), *Picea* (28.3%), and *Abies* (7.9%), along with *Tsuga* (2.9%), and hardwood species (4%). In addition, relict pollen was sporadically found. The pollen content of Ericaceae (4.5%) and the amount of moss spore are slightly increased.

The lower Urda Beds are of normal magnetic polarity. The Gauss–Matuyama boundary between the normal and reversed polarity zones (2.43 Ma) coincides with the boundary between the lower and upper Urda Beds.

The upper Urda Beds were distinguished, along with their lower unit, in the same drill sites. The interval of 316 - 376 m in Well 55 near the Urda Settlement is considered as their straotype. The beds are represented by clays with sandy and silty interlayers. Their thickness varies from 10 - 12 to 45 - 70 m.

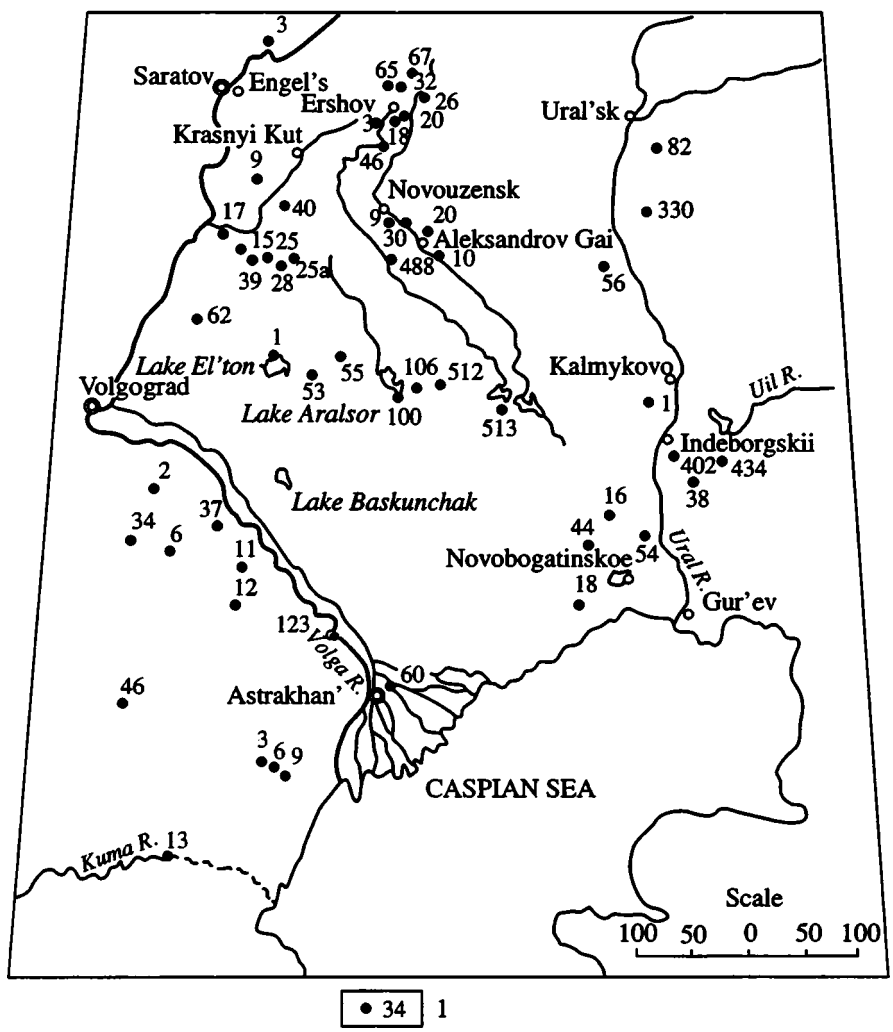

Fig. 1. Location map of drill sites.
1 – Well number and position.

The beds contain binary assemblages of marine euryhaline and brackish-water mollusks, ostracodes, and foraminifers. The total amount of genera and species is slightly lower here, and freshwater forms become more abundant among them. The bivalve species of *Avardaria* genus along with subspecies *Cerastoderma azerbaidjanicum borealis* (Fedk.) and some other forms disappeared. The bivalve mollusks *Aktschagylia venjukovi* (Andrus.), *A. ossokovi* (Andrus.), and *Kirghizella inostranzevi* (Andrus.) are the most abundant fossils of the upper Urda Beds, while *Cerastoderma novakovskii* (Andrus.), *C. abreki* (Usp.), *Aktschagylia agalarovae* (A. A.–Z.), *A. jnldsniknvi* (A. A.–Z.), and some other forms are less common. Bivalve mollusks *Dreissena polymorpha* Pall. and gastropods *Valvata piscinalis* Müll., *Planorbis planorbis* L., and others are characteristic of the peripheral regions. Ostracodes and foraminifers are represented by rare specimens of *Loxoconcha varia* Suz., *Leptocythere gubkini* Liv., and *Elphidium incertum* (Will.).

The upper Urda Beds include a pollen assemblage of the forest type similar to that from the lower Urda unit. The assemblage is peculiar because of its maximum content of *Tsuga* pollen (up to 10%), decreased amount of arborescent pollen (down to 60%), and increased content of herbaceous pollen (up to 23%). The pollen of Ericaceae predominates in the herbaceous group (9.5%), and sphagnum spores (7.2%) pre-

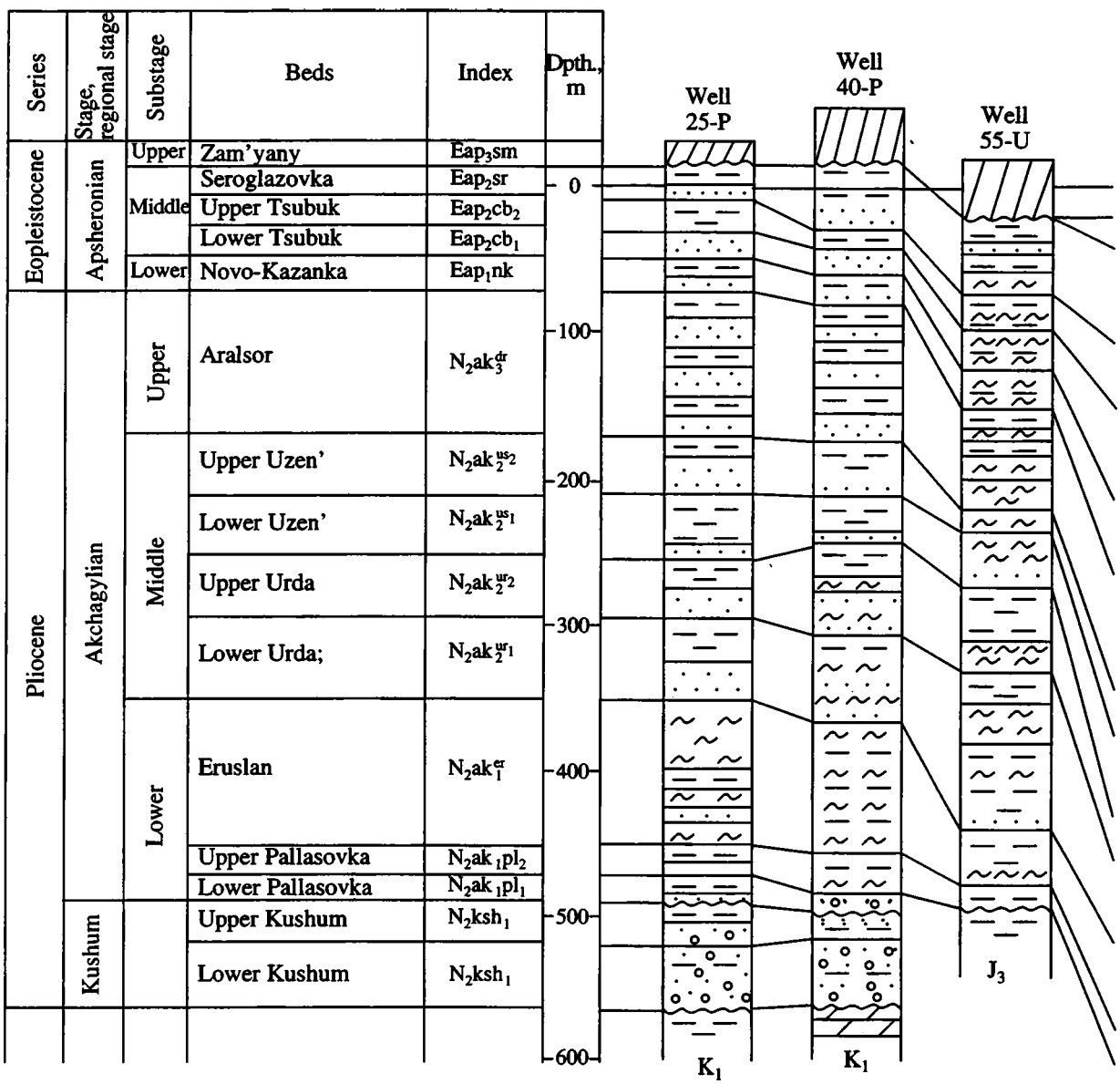

Fig. 2. Correlation scheme of the Akchagylian and Apsheronian sedimentary sequences of the Lower Volga and Northern Caspian Sea region.

1 – loams; 2 – sands; 3 – sands with gravel and pebbles; 4 – silts and siltstones; 5 – clays; 6 – marls; 7 – chalk; 8 – unconformity surface; 9 – polarity: a – normal, b – reverse.

dominate among the mosses group. The beds reveal reversed polarity of magnetization.

The Urda period corresponds to climatic warming, this phase being the optimum one in the Akchagylian epoch. The climate was temperately warm and periodically wet or dry. The temperature range was from 0 to −12°C in January and from +15 to +21°C in July during the early Urda interval. In the late Urda epoch, temperatures ranged from +3 to −2°C in January and from +19 to +22°C in July.

The lower Uzen' Beds are widespread in the Lower Volga and Northern Caspian Sea region. Their stratotype was described in Well 55 (interval 278 - 316 m), and the hypostratotype was described in Well 40 (interval 245 - 278 m) (Figs. 1 and 2). The beds are represented by clays with silty and sandy interlayers. Their thickness varies from 20 - 30 to 50 - 60 m. The beds are distinguishable due to the marine assemblage of extremely euryhaline and brackish-water mollusks and ostracodes and by the pollen assemblage of the forest (taiga) type. The molluscan assemblage includes numerous *Cerastoderma pseudoedule* (Andrus.), *Aktschagylia subcaspia* (Andrus.), *A. venjukovi* (Andrus.), and *Avicardium nikitini* (Andrus.) and rare *Cerastoderma konschini* (Andrus.), *Aktschagylia striata* (Fedk.), and other species.

Freshwater gastropods *Valvata piscinalis* Müll., associated with rare representatives of Levantine unionidas *Potomida concentricosculptus* (G.Pop.), and bivalves *Dreissena polymorpha* Pall. appear within the Ershov and Klintsovka areas. The lower Uzen' Beds contain rare ostracodes *Loxoconcha varia* Suz., *Leptocythere gubkini* (Liv.), *Prolimnocythere tenuireticulata* (Suz.), *Paracyprideis naphtarscholana* (Liv.), and other forms.

The coniferous pollen of *Pinus* (62 - 87%), *Picea* (10 - 20%), and *Tsuga* predominates in the pollen spectra

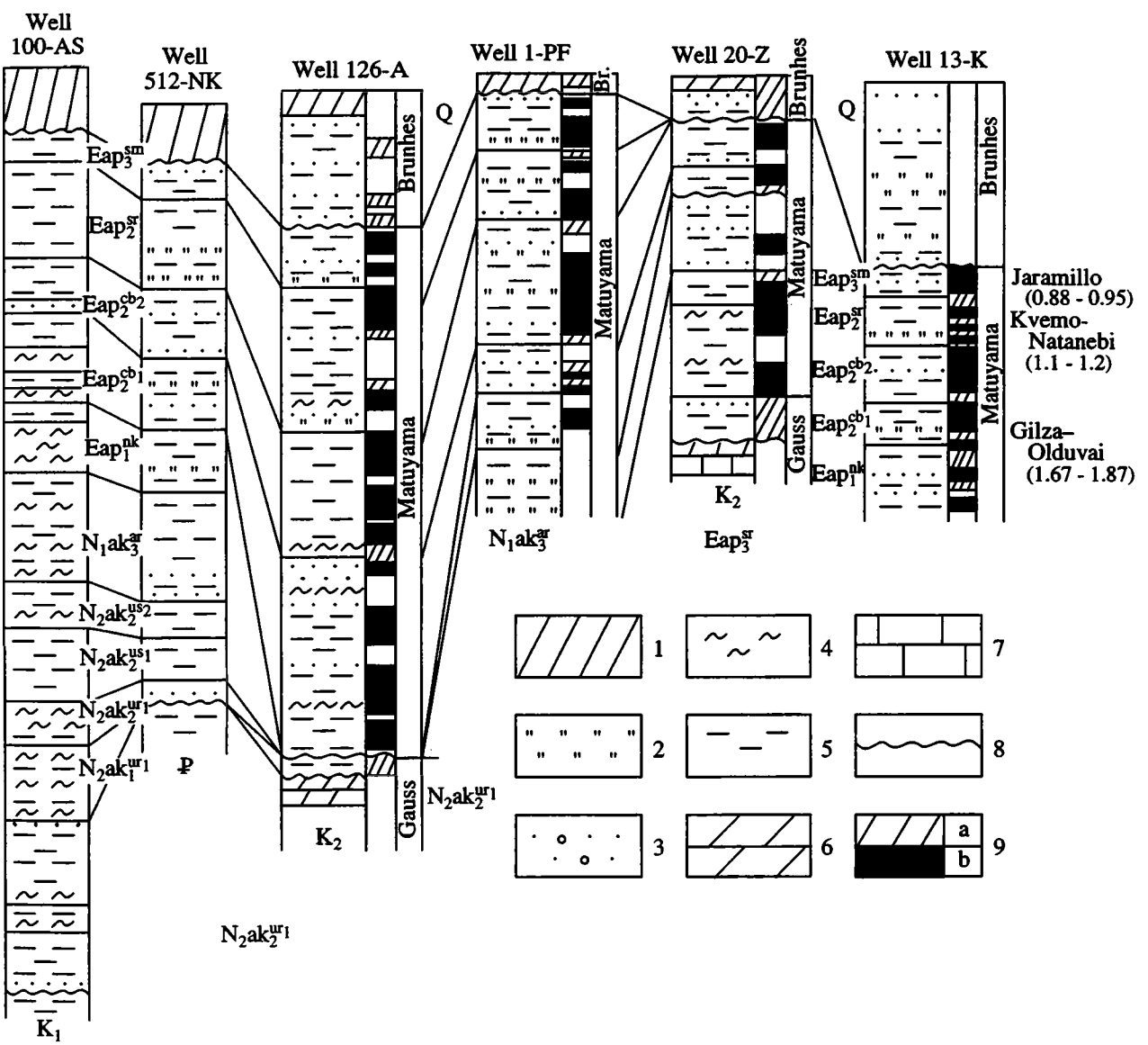

Fig. 2. (Contd.)

of the taiga type. A characteristic feature of the assemblage is the nearly complete absence of spores and hardwood and herbaceous pollen. The floral composition is indicative of spruce-pine and pine-spruce forests and a cool climate, with temperature ranges from −2 to −12°C in January and from +16 to +20°C in July.

The upper Uzen' Beds crown the middle Akchagylian sequence. Their stratotype was recognized in Well 55 (interval 260 - 278 m). The beds are represented by the intercalation of clays, silts, and sands. Their thickness varies from 10 - 15 to 50 m. They were distinguished according to their assemblage of abundant, shallow freshwater species associated with rare marine euryhaline and brackish-water mollusks and ostracodes. Bivalve mollusks include *Kirghizella inostranzevi* (Andrus.), *Aktschagylia vinjukovi* (Andrus.) (Saurbai region, Well 388 and others), *Cerastoderma vogdti* (Andrus.) (Novouzensk region, Well 20), and shallow water species *Dreissena polymorpha* Ball. (Aralsor region, Well 106 and others). Microfauna is represented by the shallow water *Cyprideis torosa* (Jones) and the fresh water *Cypria kurlaevi* Karm. and *Cytherissa lacustriformis* Popova.

The beds yielded a pollen assemblage of the steppe type. It contains mainly herbaceous specimens (58 - 92%), with xerophytes of *Chenopodiaceae, Artemisia, Artemisiidites,* and *Ephedra* taxa being the most abundant. The predominance of the steppe floral elements together with a sharp decrease of the arborescent pollen content indicate further cooling and aridization of the climate. The climate of the late Uzen' time was cool, with temperature ranges from −2 to −15°C or −16 in January and from +17 to +18°C in July. Sediments of the lower and upper Uzen' Beds have reversed polarity of magnetization.

The Aralsor Beds of late Akchagylian age crown the sequence of the Akchagylian regional stage. Their stratotype was defined in Well 100 of the Aralsor area, within the interval of 316 - 268 m (Figs. 1 and 2). They are represented by clays with silty interlayers. The thickness varies from 40 - 50 to 120 m.

The beds are distinguishable because of the impoverished assemblage of marine euryhaline and brackish-water mollusks, in which small Dreissenidae forms predominate. Freshwater forms also play an important role among them, and brackish-water cryophilic ostracodes appear in the assemblage. Small euryhaline Dreissenidae include *Dreissena distincta* Andrus., *D. polymorpha* Pall., *D. rostriformis utvensis* (Koles.), and *D. hircania* Alz. Freshwater bivalves *Pisidium amnicum* Müll. along with gastropods *Valvata piscinalis* Müll., *Viviparus* sp., and other forms are characteristic of the Aralsor Beds. In addition, the bivalve mollusks *Cerastoderma pseudoedule* (Andrus.) and *Aktschagylia subcaspia* (Andrus.), associated with gastropod *Clessiniola utvensis* Andrus., appear in the lower part of the beds. The coastal and shallow marine facies occasionally contain extremely euryhaline and freshwater species, but sometimes they yield only Dreissenidae forms (the Liman area, Well 4; Kosheli area, Well 3; Martyshi area, Well K-21; and other sites). Small Dreissenidae forms usually occur in the upper part of the unit.

Many middle Akchagylian ostracodes disappear, but *Loxoconcha eichwaldi* Liv. and *Leptocythere andrussovi* (Liv.) appear in the assemblage. Shallow water *Cyprideis torosa* (Jones) and foraminifers of *Ammonia* genera were also found there.

A pollen assemblage of the forest type (birch, spruce, and pine representatives) with abundant sphagnum spores and heather pollen is typical of the Aralsor Beds. The floral composition indicates progressive cooling and humidization of the climate, with temperature ranges from −10 to −16°C in January and from +15 to +17°C in July. The sediments in general are reversely magnetized with a normal polarity interval, which is probably correlative with the Reunion episode (2.1 Ma).

The Novo-Kazanka Beds of early Apsheronian age are widespread in the Lower Volga and Northern Caspian Sea region, close to the latitude of the Dzhanibek railway station. Their stratotype was discovered in Well 512 near the Novo-Kazanka Settlement within the interval 222 - 275 m (Figs. 1 and 2). The beds are represented by sands, silts, and clays. The total thickness varies from 40 to 130 m. The Novo-Kazanka beds were classified according to the impoverished brackish-water and freshwater fauna bearing large *Dreissena carinatocurvata* Sinz., *D. polymorpha* Pall., *D. bacuana* Andrus., and other forms dominant among mollusks. Small *Pseudocatillus* and *Monodacna* specimens are rare. A peculiar ostracode assemblage with *Paracyprideis naphtatscholana* (Liv.), *Caspiocypris candida* (Liv.), *Leptocythere andrussovi* (Liv.), and single *Cryptocyprideis bogatschovi* (Liv.) is typical of the Novo-Kazanka Beds.

The Beds contain a spore and pollen assemblage of the forest type similar to the Aralsor one. It is peculiar due to the absence of *Tsuga* pollen and a sharp increase in *Ericaceae* pollen content (up to 61%). These sediments with reversed polarity probably correspond to the Olduvai episode (1.67 - 1.87 Ma). The inferred climate was moderately cool, with temperatures ranging from −2 to −18°C in January and from +16 to +20°C in July.

The lower Tsubuk Beds of middle Apsheronian age are also widespread in the Lower Volga and Northen Caspian Sea region. Their stratotype was described in Well 123 near the Zam'yany Settlement in the Astrakhan' province, within the interval of 326 - 478 m; and their hypostratotype was described in Well 512 in the interval 180 - 251 m near the Novo-Kazanka Settlement (Figs. 1 and 2). They are represented by clays with sandy and silty interlayers. The total thickness varies from 25 to 80 m. The beds contain an impoverished and monotonous assemblage of brackish-water mollusks and ostracodes and a pollen assemblage of the forest-steppe type. Brackish-water forms of *Parapscheronia raricos-*

tata (Sjoegr.), *Pseudocatillus isseli* (Andrus.), and *Didacnoides bakuanus* (Andrus.) are abundant (the Urda area, Well 55; the Aralsor area, Well 106; and other localities). Species of the euryhaline *Dreissena, Adacna, Monodacna, Micromelania*, and other genera are characteristic of the shallow water facies, while *Pseudocatillus isseli* (Andrus.) and *P. catilloides* (Andrus.) are rare.

The lower Tsubuk Beds also include brackish-water species: *Caspiolla acronasuta* (Liv.), *Caspiocypris candida* (Liv.), *Paracyprideis naphtatscholana* (Liv.), *Leptocythere andrussovi* (Liv.), *Loxoconcha eichwaldi* Liv., *Mediocythereideis*, and others.

A pollen assemblage of the forest-steppe type characterizes the lower Tsubuk Beds, bearing mainly *Pinus* and *Betula* pollen (5.6%) of the arborescent plant group. Pollen of the broad-leaved species (1.9%) includes specimens of *Ulmus, Tilia, Corylus, Quercus, Carpinus, Acer*, and *Fraxinus*. Single grains of relict species appear in the assemblage again, thus indicating slight climatic warming. The amount of Ericaceae pollen sharply decreases (down to 6.6%). The amount of steppe forms (Chenopodiaceae, *Artemisia, Ephedra*) increases (up to 36%), as does the amount of herbs (4.9%) and cereals (4.3%). Mosses (10.4%) and ferns (5%) predominate among the spores.

The upper Tsubuk Beds occur in the distribution areas of the lower Tsubuk Beds. Their stratotype was described in the interval of 246 - 326 m of Well 123. They consist of clays with interlayers of sands and sandstones. Their thickness varies from 50 to 100 m. The beds are characterized by a rich and diverse mollusk and ostracode assemblage, which includes nearly all genera and species known in the Apsheronian deposits of the Lower Volga and Northern Caspian Sea region. Such species as *Parapscherona raricostata* (Sjoegr.), *P. eurydesma* (Andrus.), *Pseudocatillus isseli* (Andrus.), *Monodacna laevigata* Andrus., *M. sjoegreni* Andrus., *Hyrcania intermedia* (Eichw.), and others predominate in the sections of the central Caspian Sea region. Similar molluscan assemblages were found in the peripheral parts (Novouzen' area, Parfil', and Kumskii settlements, and other localities).

The rich ostracode assemblage of the beds contains abundant forms of thermophilic genera *Leptocythere, Loxoconcha, Aurila*, and rare eurythermal species *Caspiolla acronasuta* (Liv.), *Paracyprideis naphtatscholana* (Liv.), *Criptocyprideis bogatschovi* (Liv.), and others. The amount of freshwater species increases in the upper part of the section.

A pollen assemblage of the forest-steppe type similar to that of the lower Tsubuk Beds was found in the upper unit. It is distinct due to the higher abundance of steppe plant pollen (up to 54.7%) and the greater diversity of herbs (5.9%). At the same time, the amount of Ericaceae (1.2%) and small-leaved *Betula* (3%) pollen decreases along with a decrease in spore abundance. The lower and upper Tsubuk Beds show reversed polarity (the Matyama orthozone). The climate of the Tsubuk age was moderately warm, with the temperature ranging from −4 to −10°C in January and from +22 to +24°C in July. The end of the Tsubuk time was a period of aridization, as is evident from the increased amount of Chenopodiaceae pollen and the decreased diversity of arborescent pollen.

Their stratotype was described in Well 55 (interval of 41 - 98 m) near the Urda Settlement (Figs. 1 and 2). The beds are represented by clays with sandy and silty interlayers. Their thickness varies from 60 - 70 to 100 - 140 m. The beds contain several assemblages of brackish- and freshwater mollusks and ostracodes. Many middle Apsheronian species became extinct at this time, being replaced by bivalves of *Dreissenidae* and by gatropods of the *Valvatidae* families. Still, the *Parapscheronia eurydesma* (Andrus.), *Monodacna laevigata* Andrus., *Pseudocatillus dubius* Andrus., and other forms of mollusks along with species of the *Caspiocypris, Caspiolla*, and other ostracode genera continued to occur.

A pollen assemblage of the steppe type with a dominant herbaceous (67 - 84%) group characterizes the beds. *Chenopodiaceae* pollen predominates. The pollen of herbs and Gramincea is diverse. The amount of arborescent pollen and spores is negligible. In general, the Seroglazovka Beds reveal reversed polarity, interrupted by two intervals resembling the Kvemo-Natanebi (about 1.1 Ma) and Jaramillo (0.89 - 0.95 Ma) episodes. The climate during the Seroglazovka time was still moderately warm, tending to become more arid and continental. The temperature ranged from −2 to −16°C in January and from +14 to +20°C in July. This corresponds to the presence of both the thermophilic (*Corbicula, Valvata, Unio*) and moderately cryophilic (*Cypria, Iliocypris*) forms.

The Zam'yany Beds of the Eopleistocene age crown the sequence of the Apsheronian regional stage. Their stratotype was described in Well 123, within the interval of 110 - 140 m. The beds are represented by clays with sandy and silty interlayers. Their thickness varies from 30 to 150 m. A monotonous assemblage of brackish- to freshwater mollusks and ostracodes distinguished the beds. The brackish- and freshwater forms of *Dreissena, Micromelania, Caspia*, and *Valvata* genera predominate among the mollusks (Aralsor area, wells 100, 106; Novo-Kazanka area, Well 512; the Ural River, wells 2, 13, 14; and others). The bivalve mollusk *Apsheronia propinqua* Eichw., characteristic of the upper Apsheronian deposits in the Lower Volga and Northern Caspian Sea region, was found in sections of the Astrakhan' area (wells 60, 123, and others). The assemblage also includes rare specimens of *Leptocythere reniformis* (Liv.), *Caspiolla acronasuta* (Liv.), *C. gracilis* (Liv.), *Caspiocypris filona* (Liv.), *C. mandelstami* (Liv.), and *Leptocythere bakuana* (Liv.) along with the first appearance of cryophilic stenothermal *Cytherissa lacustris* Sars.

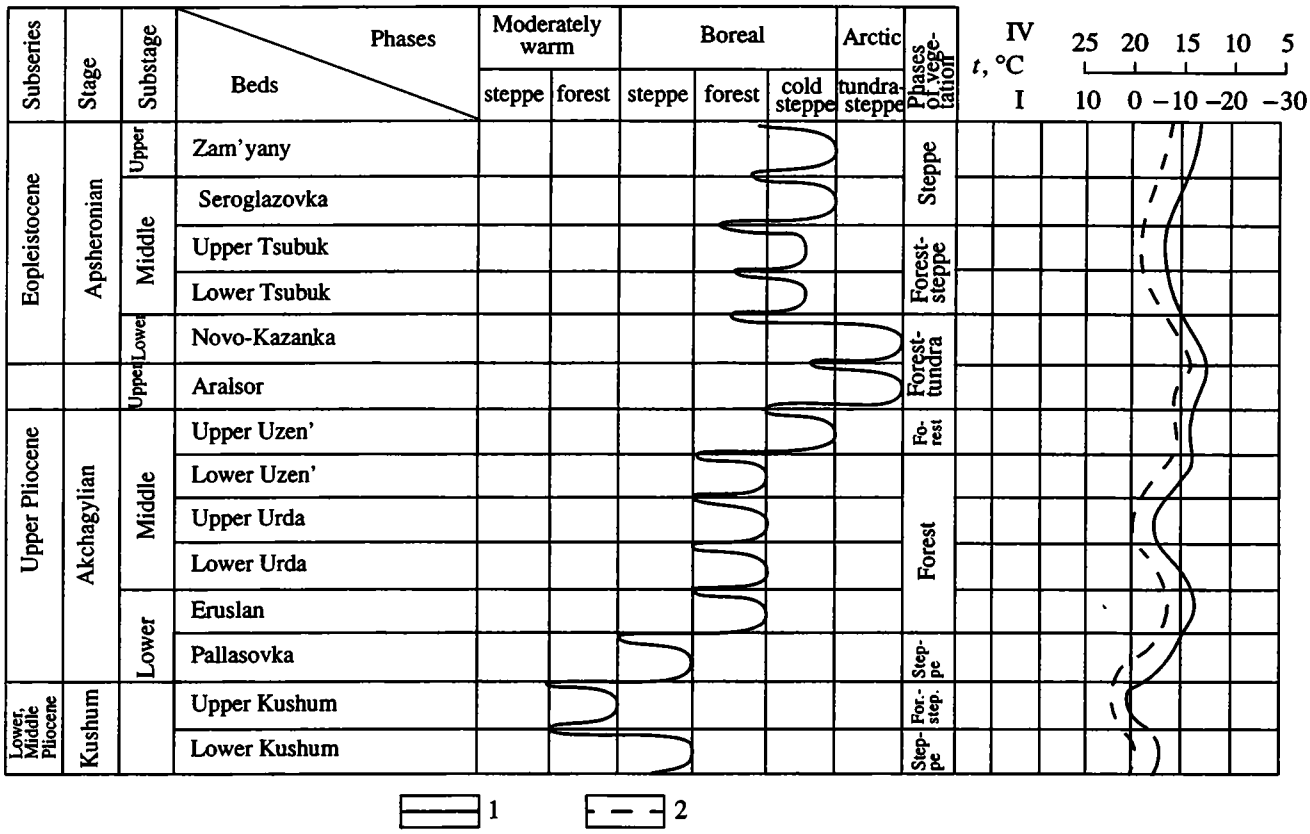

Fig. 3. Curves of the vegetation cyclic evolution and temperature fluctuations in the Lower Volga and Northern Caspian Sea region during Pliocene and Eopleistocene.
1 – January temperature; 2 – July temperature.

The Zam'yany Beds contain a spore and pollen assemblage of the steppe type. Small, uniform pollen of the *Chenopodiaceae* is dominant in it (50 - 98%). Pollen of *Artemisia* and *Gramineac* is also present in the assemblage, but the amount of herbs decreases. Spores and arborescent pollen are rare. Floral composition indicates that further aridization of the climate resulted in the impoverishment of flora and the nearly complete disappearence of arborescent elements. The climate remained cool, with temperature ranges from −2 to −16°C in January and from +14 to +21°C in July.

In general, the climate of the late Pliocene and Eopleistocene was boreal, with progressive cooling by the end of the Eopleistocene. It was warmer and more humid in the Akchagylian time than during the Apsheronian interval. The presence of fir, tsuga, spruce, and representatives of the *Taxodiaceae* family in the Akchagylian paleoflora proves this statement (Fig. 3).

Reviewer M.N. Alekseev

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Geochronology and Stratigraphy of the Holocene Lacustrine-Bog Deposits in Northern and Central Kazakhstan

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Abstract – A detailed stratigraphic subdivision of five sections of lacustrine-bog deposits in northern and central Kazakhstan, based on comprehensive palynological data and radiocarbon datings, is obtained for this area for the first time. The time scale of climatic cycles and pollen zones covers the period from the Allerød until the present and shows a satisfactory correlation between the vegetal and climatic phases and the Blytt–Sernander periodicity scale for the Holocene with Khotinskii's modification (1977, 1989). The two climatic optima established in the Holocene are equally important: the early Atlantic optimum (8 to 7 ka) and the middle sub-Atlantic optimum (1.8 to 0.8 ka). The beginning of sapropel accumulation on a large scale in the freshwater lakes of the steppe area in Kazakhstan (about 10 ka) is considered synchronous with a similar phenomenon in many lakes of the temperate belt of northern Eurasia. This phase of accumulation, therefore, can be used as a stratigraphic boundary between late and post-glacial time as well as for distant geographical correlations.

The study of the Holocene evolution of the environment in the steppe area of northern and central Kazakhstan is important for the solution of many problems in paleogeography, stratigraphy, and correlation. In this respect, radiocarbon datings and palynological studies of post-glacial deposits offer the best opportunities. Radiocarbon datings allow us to reach the current maximum accuracy when defining the age of Holocene and upper Pleistocene deposits and to correlate paleogeographical schemes in any part of the Earth. In addition, the palynological method provides an important link between radiometric and paleogeographical data (Khotinskii, 1977). Pollen diagrams comprise an extensive amount of information on the character of vegetation in the past, which is a sensitive indicator of variations in the environment and, primarily, in climatic conditions.

The peat and sapropel deposits in the fresh-water lakes and peat bogs in the territory of Kazakhstan provide ample information for climatostratigraphic and paleogeographic research. These concordantly bedded sediments contain, along with other biogenic remains, abundant pollen and spores of higher plants. Less favorable for research are the alluvial sediments of the flood-plains and terraces because of the low content of organic matter, which hinders their radiocarbon datings, and the frequent redeposition of pollen and spores (Chupina, 1974).

The palynological method was first applied in order to study the peat sections of the Borovoe Reservation steppe area in the Kokchetav region of Kazakhstan (Zharkova, 1930); however, only the pollen of arboreal plants were identified. The discovery of pollen of broad-leaved oak, lime, and elm in the lower horizons of the studied sections was sensational, because their provinces are currently located much further north, beyond the Kazakhstan border (Gorchakovskii, 1987). According to Zharkova (1930), the peat bogs appeared at the beginning of the Holocene, when the climate became drier due to post-glacial warming. With further research, Zharkova somewhat modified her views on the history of the landscapes in this region and the time when the peat started its accumulation. Palynological study of the peat bog on the bank of the Svetloe Lake revealed three phases in the evolution of the landscapes and climate since the initiation of peat formation (Zharkova, 1967). The lower layers (2.7 to 3.5 m) represented a phase of "a more humid climate, when the lake was surrounded mainly by birch and pine forests; the spectra also contain pollen of the broad-leaved and dark-coniferous varieties." The middle layers of the peat bog (1.5 to 2.7 m) lacked the pollen of dark-coniferous and broad-leaved trees, but their content of alder was somewhat higher and indicated "a xerothermic phase with a drier climate." The pines absolutely dominated in the spectra of the upper part of the peat (0.0 - 1.5 m), which accumulated during "a phase with a modern climate, which is slightly more humid and colder." Zharkova (1967) dated the beginning of peat accumulation in the Kokchetav low-mountain area as the middle Holocene.

In the 1940s, prospecting for peat was greatly encouraged in northern Kazakhstan and, consequently, interest in the stratigraphy of peat deposits was high.

The results from the palynological study by Krupenikov (1941) of a peat deposit 1.2 m thick in the Naurzum Reservation of the Kustanai region implied more humid time periods in the Holocene, and he referred the beginning of peat formation to the boreal period of the Holocene. An important fact is that the pine pollen in the spectra from the lower peat layers constituted only a small percentage of the total amount of arboreal pollen, whereas it dominated in the upper part of the section. It is also essential to note that the base of the Borovoe peats also contained a much lower amount of pine pollen than the recent peat horizon.

Lavrov (1945, 1946), having studied the peat deposits in the Sukhoe Zaimishche (1.2 m thick) and Shaitomar (1.4 m thick) localities near Kustanai, noted the presence in the pollen spectra of many arboreal plants: pines, birches, spruces, firs, Siberian pines, alders, and limes. Lavrov attributed the diversity of pollen to its introduction from distant sources. He also noted a lower content of pine pollen in the lower parts of the sections and individual lime pollen grains. His first age determination of the peat bogs was 2 ka (Lavrov, 1945); but in a later publication (Lavrov, 1948), he brought their age to 4500 yr.

In his review of the paleogeography of USSR territory in the Holocene, Neishtadt (1957) concluded that the peat bogs of Kazakhstan were formed under the conditions of a steppe landscape with oases of pine and birch forests. He considered the presence of pollen from broad-leaved plants in the lower parts of some of the sections as a regular feature of specific botanic, geographic, and palynostratigraphic significance. He believed that only radiocarbon datings, detailed sampling for palynological studies, and analysis of grass pollen and spores can unequivocally establish whether peat layers with pollen of broad-leaved plants belong to the Holocene or whether they are the remnants of older epochs (Neishtadt, 1957). He also pointed out another essential drawback in the paleogeographical schemes of Kazakhstan compiled before 1950: the pollen of segebrushes, the inevitable members of steppe coenoses, was mistaken for that of willows. As a result, changes in the content of the latter were mistaken for the alternation of dry and humid climatic conditions.

More recently, studies of the late Pleistocene and Holocene in northern and central Kazakhstan were focused on the alluvial sediments of the terraces and flood-plains of the Tobol, Ishim, and Irtysh rivers and their tributaries (Klapchuk, 1968; Chupina, 1974; and others).

The results of long-standing paleogeographical studies of late post-glacial time in the steppe area of Kazakhstan, mainly derived from palynological data on flood-plain sections, revealed the following pattern of landscape-climatic changes (Aubekerov *et al.*, 1989; Nurmambetov *et al.*, 1989). These authors referred the pollen spectra with an absolute dominance of Chenopodiaceae, *Ephedra*, Compositae, and sagebrushes to a phase of dry and cold climate during the late Dryas.

The pollen assemblages referred to the Holocene pre-boreal and boreal periods contained a slightly larger amount of pollen of arboreal plants and a greater diversity of their species. Among grasses, the number of mesophilic elements increased; spores also appeared as evidence of a warmer and more humid environment than that of the previous phase. As stated by Aubekerov *et al.* (1989), the post-glacial period and the beginning of the Holocene had a drier, colder, and definitely more continental climate than the recent period.

A larger share of mesophilic vegetation is typical of the Atlantic time. The spore-pollen spectra contained more abundant and diverse pollen of arboreal plants; of particular interest was the presence of pollen of broad-leaved limes and maples, as well as that of alders and spruces. This vegetation indicated the warmest and the most humid climatic phase of the Holocene. The mid-Atlantic time is assumed to be the Holocene climatic optimum (Aubekerov *et al.*, 1989).

The sub-boreal period is characterized by pollen assemblages whose composition is indicative of a drier climate than that of the mid-Atlantic. Pines dominated in the arboreal vegetation, whereas the steppe assemblages were more abundant in the herbal part of the spectra.

The type of spectra that Abukerov *et al.* (1989) referred to the Holocene sub-Atlantic time indicated a landscape-climatic environment similar to the recent one. As the assemblages of herbals were replaced by the Poaceae-sagebrush vegetation of the steppes, the share of Chenopodiaceae-sagebrush species increased due to the drier and cooler climate.

While acknowledging the progress that has been made in the study of the paleogeography of Kazakhstan, we are bound to point out its rather sketchy reconstruction. In the absence of radiocarbon datings, researchers have to use correlation with distant geographic areas when correlating identified palynozones with the time scale. This hampers the solution of many debatable problems and the comparison of data of different researchers.

Since 1989, we have conducted radiocarbon datings in order to more accurately define the age of the pollen zones of the late glacial and the Holocene in northern and central Kazakhstan. The lacustrine – bog deposits of mainly sapropel and peat were considered to be the most advantageous for this study. The samples for palynological analysis were obtained layer-by-layer with a peat bore every 10 cm, sometimes every 5 cm. The samples of peat and sapropel for radiocarbon datings and botanical analysis were also collected with sufficient frequency. Detailed spore-pollen diagrams and 35 radiocarbon datings are now available for five sections (Tarasov, 1991; Tarasov, 1992); their localities are shown in Fig. 1. The correlation between the acquired data is satisfactory and represents a complete

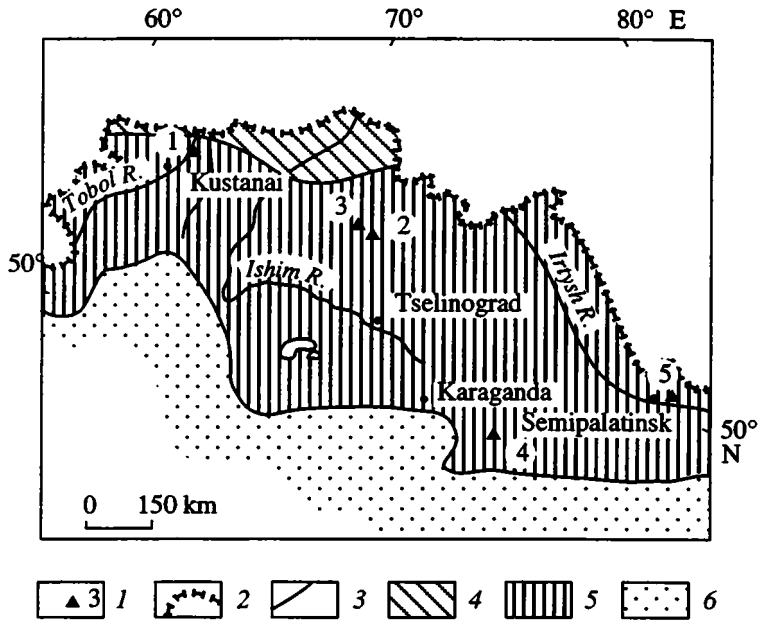

Fig. 1. Localities of studied sections of lacustrine and lacustrine-bog deposits.
1 – locality of sections (1 – Mokhovoe, 2 – Kotyrkol', 3 – Karas'e, 4 – Pashennoe, 5 – Ozerki); *2* – the boundary of Kazakhstan; *3* – the boundaries of natural zones; *4* – forest-steppe; *5* – steppe; *6* – semidesert.

sequence of vegetal and climatic evolution of the area from the Allerød until the present, with some peculiarities in the spore-pollen spectra caused by local geological-geomorphological and microclimatic features.

The most typical of all the sections studied is represented by the bottom deposit of the Pashennoe Lake (49°22′ N, 75°24′ E; 871 m above sea level). The lake is situated in the south of the steppe zone within the Karkaralinsk mountain region. The lake covers about 30 hectares and is surrounded on three sides by granite hills about 1000 m high. On the southern side, where there are no mountains, the Kuruuzek Brook issues from the lake. The larger part of the lake's surface is covered with floating bogs, which build up rather large separate islets; their surface is springy, hummocky, and overgrown with bushes of willows, birches, and aspens; the moss-grass layer is thin.

The narrow shore band around the lake is covered with various aquatic and swamp vegetation (sedges, equisetums, reeds) and occasional patches of accumulating peat. A birch forest grows over the peat bog in the northern part of the lake. The Karkaralinsk low-mountain upland is one of the southernmost areas in the steppe zone of Kazakhstan with pine forests and an admixture of boreal relics (Gorchakovskii, 1987). The crests of ridges and the steep, eroded slopes of the granite hills are covered by thin pine forests with rare *Betula rerrucosa*. The density of trees and the diversity of plants is greater at the base of the slopes. Different vegetal varieties of *Caragana–Festuca–Stipa* steppes with *Ephedra* cover the surrounding hillocky plains.

The recent climate of the region is continental. In the Karkaralinsk area, the average temperatures in July are +18 - 19°C, in January they are –14 - 15°C, and the average annual temperature is about +2°C. The average amount of precipitation is 300 mm/y (*Agroklimaticheskii Spravochnik...*, 1962).

We drilled a bore-hole in one of the floating bogs, which was built up of slightly decomposed sedge peat. The bore penetrated the 2.25-m water layer and entered the following strata of limnetic deposits (Fig. 2):

From 0.0 to 2.5 m: clayey, greyish brown sapropel with inclusions of vegetal remains; the lower part contains shells of the mollusks *Pisidium* (*Eupisidium*) *amnicum*, *Radix lagotis*, and *Gyraulus gredleri*; the sapropel is composed mainly of the remains of diatoms; in the lowermost 30 cm layer, chlorococcal algae and the remains of higher plants dominate;

From 2.5 to 3.3 m: limnetic blue-grey clays with inclusions of granite gruss and shells of freshwater mollusks (the remains of *Planorbis* and *Limnaca* supplement those mentioned above); the share of organic matter in the upper part of the clays increases noticeably.

The abundance of granite gruss in the lower clay layer is apparently a result of the proximity of an indigenous bed rock.

Sediment accumulation in the lake started about 12 ka in the late Pleistocene. This date of 11 710 ± 100 yr. (TO-3026) was established by radiocarbon dating of the base of the bore-hole drilled in the Ozerki bogs near Semipalatinsk (Fig. 1). The spore-pollen spectra, recovered from the base of the Ozerki section, are in

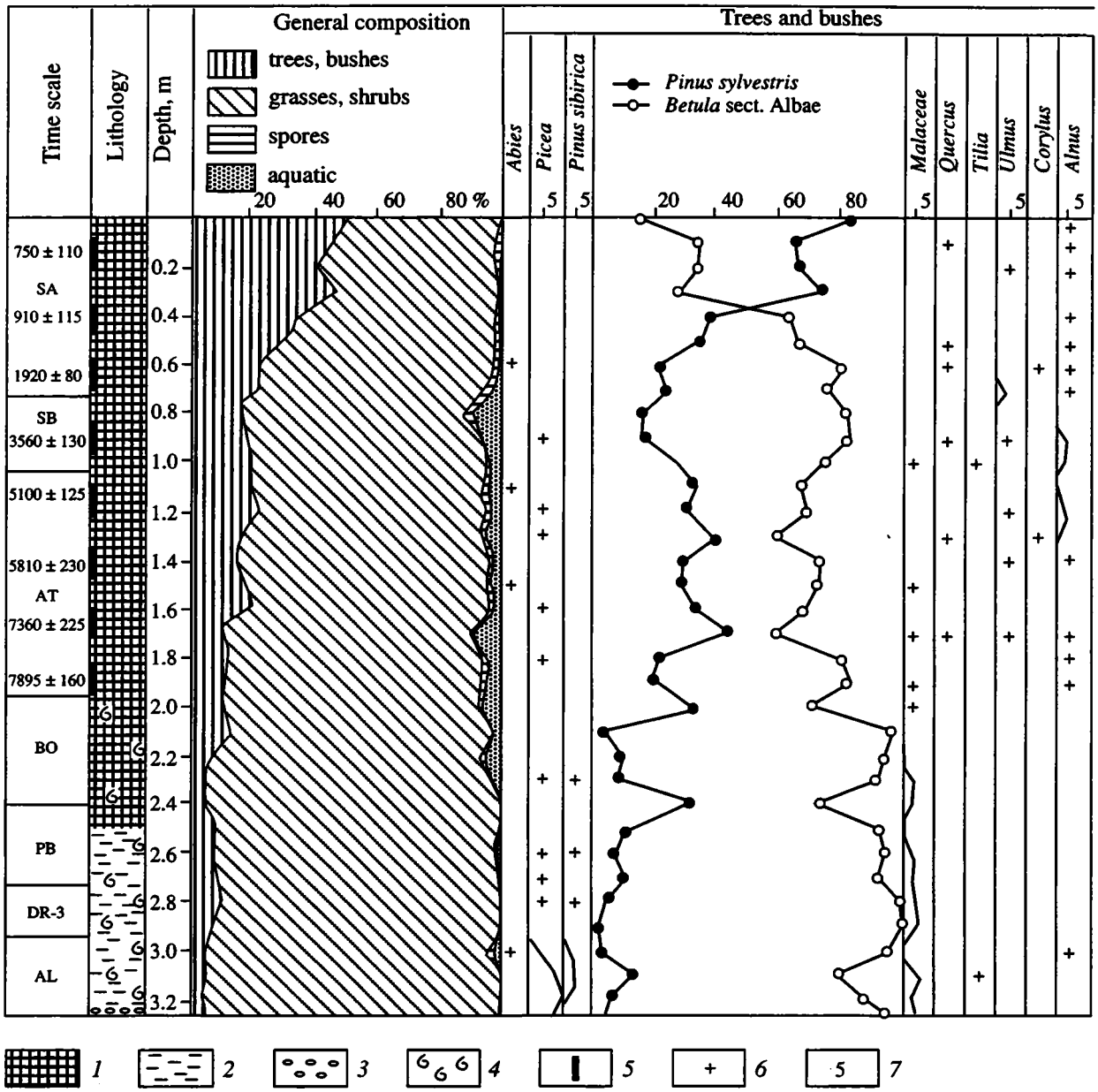

Fig. 2. The spore-pollen diagram of the section of the bottom sediments in the Pashennoe Lake.

1 – organic-clay sapropel; *2* – clay; *3* – gruss; *4* – shells of mollusks; *5* – intervals sampled for radiocarbon datings; *6* – pollen content less than 2% in the intragroup ratios; *7* – absolute amount of pollen grains and spores in cases when intragroup ratios were not calculated (less than 50 specimens in a group). Climatic periods of late glacial time–Holocene (after the Blytt–Sernander scheme with Khotinskii's modification): AL – Allerød, DR-3 – Younger Dryas; PB – pre-boreal; BO – boreal; AT – Atlantic; SB – sub-boreal; SA – sub-Atlantic.

complete agreement with similar spectra from the bottom deposits of the Pashennoe Lake.

In the clay layer at a depth of 2.95 - 3.3 m (Fig. 2), the pollen assemblages dated as the Allerød contain pollen of diverse arboreal and bush plants, with a prevalence of grasses and undershrubs. Noteworthy is the abundance of pollen of spruces, sea buckthorn, and *Lonicera*; up-section, their pollen appears sporadically. The composition of the pollen spectra indicates that the vegetal cover consisted mainly of Poaceae-sagebrush and sagebrush steppe plants with Chenapodiaceae and grass assemblages. Thin forests with birches, spruces, and Siberian pines grew only in the most favorable conditions of the river and lake valleys. The underbrush varieties include willows and *Lonicera*; the sea buckthorn grew on well-lighted slopes with thin turfs. The climate during that period was colder and more humid than it is at present due to lower summer temper-

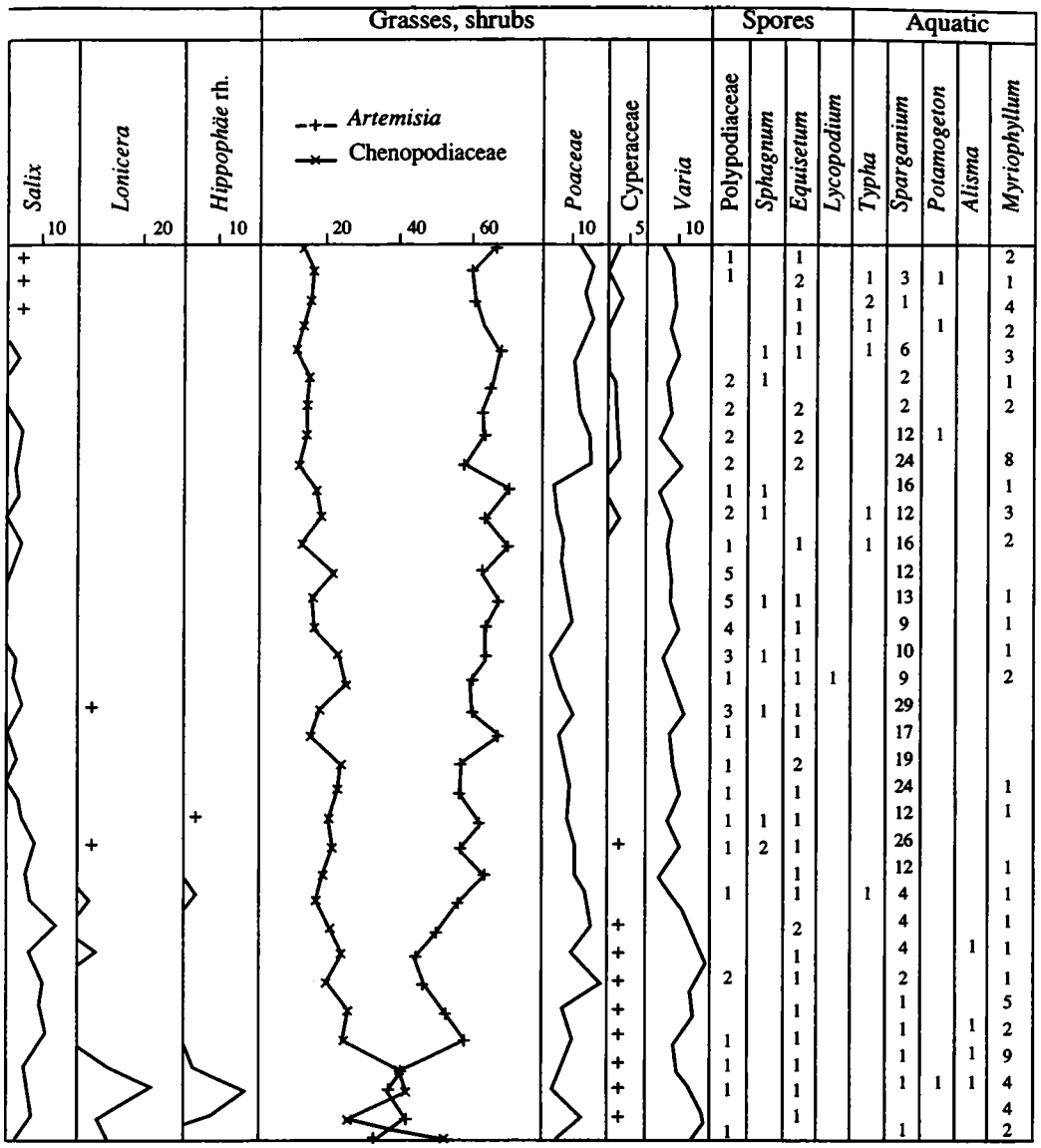

Fig. 2. (Contd.)

atures and evaporation. The lake was in a transgression phase, as evidenced by the composition of the pollen of macrophytes and by the rather diverse mollusk assemblages.

The transition to the late Dryas, 11 ka, is clearly reflected in the composition of the pollen spectra by the sharp reduction of coniferous; the disappearance of spruces, Siberian pines, sea buckthorn, and *Lonicera*; and the somewhat higher absolute content of birch pollen. In the grass-undershrub group, the share of sagebrush increases, whereas that of herbs decreases. We attribute these changes to cooling and a more continental and drier climate, which caused a reduction in forest assemblages (except the birch and willow groves on the banks of the rivers and lakes) and xerophytization of the steppe vegetation in northern and central Kazakhstan. The landscape-climatic environment of the late Dryas (11 000 to 10 200 yr) probably had features in common with that of the Sartan Ice Age in western Siberia, when the tundra-steppe landscape developed on the territory of northern Kazakhstan (Aubekerov *et al.*, 1989). The changes in the pollen composition of macrophytes (Fig. 2) and the appearance of bog species also imply dessication of the climate and regression of the Pashennoe Lake.

The interpolation of radiocarbon datings throughout the section using deposition rates allows us to confidently refer the 2.4 - 2.75 m layer to the pre-boreal period of the Holocene, i.e., about 10 200 to 9200 yr. The pollen assemblages of that time contain more arboreal plants, a smaller share of sagebrushes, and a greater amount of Poaceae and grasses. The climate was warmer and less continental. Higher temperatures

caused active eutrophication of the lake, thus starting the clayey sapropel sedimentation.

The change in the type of sediments in the lake is, in fact, an important stratigraphic boundary between late glacial time and the Holocene. In the Ozerki section, the time of the 10-cm-thick peat sapropel layer above the limnetic clay-sapropel boundary is 9280 ± 380 yr. (MGU-1320). The radiocarbon dating of the sapropel base (9610 ± 100 yr) in the grey-blue clays of the Borovoe Lake on the Kokchetav Upland, which is in close proximity to the Karas'e section (Berdovskaya, 1989) (Fig. 1), confirms that the process was by no means local. An analysis of the literature concerning other regions in northern Eurasia (Khotinskii, 1977; Andreev *et al.*, 1989; and others) implies that the beginning of sapropel sedimentation in lakes may serve as a basis for stratigraphic and paleogeographic correlations with distant areas.

We determined the pre-boreal–boreal boundary in the deposits of the Pashennoe Lake to be 2.4 m, based on a rather short-term increase in the share of pines in the pollen spectra along with a general decline in the content of arboreal-bush pollen. The boreal pollen assemblages (1.95 - 2.4 m) are characterized by the most abundant herbal pollen and an almost complete absence of pines. The deposits of the Ozerki section represent a striking example in this respect, because some of their samples lack even single grains of pine pollen. The composition of the spectra suggests that 8 - 9 ka, in northern and central Kazakhstan, the steppes spread over larger territories because of a higher aridity of the climate. While the province of the pines diminished, birch occupied the drying banks of the regressing lakes.

The boundary between the boreal and Atlantic periods is drawn at 1.95 m, based on changes in the composition of the arboreal-bush pollen spectra. The time of the boundary is reliably determined as 7895 ± 160 yr (IGAN-1278) by radiocarbon dating of the 1.85 to 1.95 m layer. The common features of the pollen spectra of the Atlantic period are a larger share of pines; the appearance of alders; and, later, the presence of elms, oaks, and limes. In addition, the amount of pollen of steppe vegetation is reduced, undoubtedly due to the more humid and less continental climate. An analysis of the modern climatic environment in the southern and southeastern confines of the oak, elm, and lime provinces indicates (Tarasov, 1992) that during those periods, in comparison with the present, the amount of atmospheric precipitation should have been 100 - 120 mm/y more, the average temperature in July, 1 - 2°C lower, and the temperature in January, 1 - 1.5°C higher to allow the broad-leaved plants to move into the forest-steppe and steppe zones of Kazakhstan.

However, an analysis of pollen spectra suggests repeated changes of climate during the Atlantic time in the Holocene and their effect on the landscape environment. The spore-pollen data imply three pluvial phases and three phases of a relative dessication of the climate. We determined the age of these phases by radiocarbon datings from the 1.05 - 1.95 m layer, which we refer to the Atlantic period: 7895 ± 160 (IGAN-1278), 7360 ± 225 (IGAN-1187), 5810 ± 230 (IGAN-1279), and 5100 ± 125 yr (IGAN-1185). The most favorable climatic conditions were about 7.5, 5.5, and 4.8 ka; the early Atlantic phase (8 to 7 ka) corresponds to the Holocene Atlantic optimum.

We draw the Atlantic – sub-boreal boundary at the 1.05 m level on the basis of a sudden decline in the pine pollen concent, the disappearance of pollen of broad-leaved plants, and a general reduction of arboreal pollen in the spectra. The interpolation of radiocarbon datings of 5100 ± 125 yr (IGAN-1185) and 3560 ± 130 yr (IGAN-1280) dates the boundary at about 4600 yr. The composition of the pollen assemblages of the sub-boreal period suggests that at that time the environment was unfavorable for arboreal plants. The territory of northern and central Kazakhstan 2.5 to 4.5 ka was increasingly overrun by steppes. Among grasses, the sagebrush groups, with contributions from Poaceae and herbal assemblages, dominated. The forest areas were smaller than they had been in the previous period, and broad-leaved plants could survive only in the most advantageous localities. The aridization of the climate in the early sub-boreal time lowered the water level in the lakes and encouraged swamping and peat accumulation in them. Radiocarbon datings of the base of the peat layer are 2760 ± 130 yr (MGU-1313) in the Mokhovoe peat bog and 3485 ± 110 yr (MGU-1263) in the Kotyrkol' peat bog; these data imply that most of the peat bogs in the steppe zone of northern and central Kazakhstan started their formation about 3500 yr. The pollen spectra from the Karas'e, Kotyrkol', and Mokhovoe sections (Fig. 1) are correlative with those from the Pashennoe Lake. In northern Kazakhstan, the indicated variations were considerably less only on the Kokchetav Upland due to a more favorable microclimate; it is as if atmospheric precipitation was "attracted" by the low Kokchetav Upland (*Klimat Kazakhstana*, 1959).

The beginning of the sub-Atlantic time of the Holocene is recorded by an abrupt increase in pine pollen content, which changed the relationship between the arboreal and herbal parts of the spore-pollen spectra. The age of this boundary (about 2500 yr) was established by interpolation of radiocarbon datings of 3560 ± 130 yr (IGAN-1280) and 1920 ± 80 yr (IGAN-1183) (Fig. 2). The sub-Atlantic was, therefore, a time of overall dissemination of pines within the steppe zone of Kazakhstan; they are currently found growing in the following locations: pine forest oases stretched along river valleys on the sandy terraces of the Tobol, Ubagan, Ishim, and Irtysh rivers; on the granite and metamorphic rocks of low mountains; and on the Kokchetav Upland (Gribanov, 1960). Judging from the climate of this period, we may state that on this territory during the Holocene the most advantageous combination of thermal conditions and precipitation occurred in the

mid-sub-Atlantic phase (about 1800 to 800 yr). This conclusion is supported by the composition of the palynospectra, the much higher sedimentation rate in all of the lacustrine and bog sequences studied, and the radiocarbon datings of the sub-Atlantic deposits from the 0.0 - 0.75 m layer of the Pashennoe Lake section: 1920 ± 80 yr (IGAN-1183), 910 ± 115 yr (IGAN-1281), and 750 ± 110 yr (IGAN-1188). Because of the small thickness of the upper sub-Atlantic sediments, we could not obtain their radiocarbon datings or detailed paleobotanical characteristics. We can only state that the pollen assemblages of the late sub-Atlantic phase contain a smaller amount of the pollen of the pines and none of the pollen of the broad-leaved plants. This probably relates to climatic variations and anthropogenic impacts on the steppe zone during the last millenium.

CONCLUSION

The results of palynological research and radiocarbon datings of the lacustrine-bog deposits allowed us to correlate the phases in the evolution of vegetation and climate in northern and central Kazakhstan with the modified Blytt–Sernander time scale (Khotinskii, 1989), which is commonly used in the study of the Holocene paleogeography of Eurasia.

The history of the landscape climatic environments of the region in late glacial time was studied for the first time; as a result, we endeavoured to reliably define the Allerød Interstadial and the late Dryas cooling. The vegetation developed from the almost woodless, "cold" steppes of late glacial time to the Atlantic forest-steppe landscapes, covered with herbal–cereals–sagebrush and herbal–cereals steppes and groves of pines, birches, and broad-leaved plants migrating into northern Kazakhstan from the southern Urals and western Siberia. The pine forests spread over the largest areas during the Holocene sub-Atlantic period. The vegetation in the final phase of the Holocene acquired the appearance of recent flora, and the forests lost their broad-leaved components.

The transition from the cold climate of the late glacial to the much warmer Holocene is established by palynological data, changes in the lithology of lacustrine deposits, and the beginning of an excessive accumulation of sapropel in fresh-water basins.

In the studied area, the temperature-precipitation ratio varied from a highly continental cold dry climate in the late Dryas, after the milder environment of the Allerød Interstadial, to a much warmer and rather dry climate of the pre-boreal and boreal periods in the Holocene. The milder continental environment and, as a consequence, the higher humidity of the climate typical of the Atlantic time reached their optimum about 7.5 ka. This was followed by a slight trend towards aridization, particularly in the sub-boreal time, which initiated the swamping of lakes and peat accumulation. The second well-developed Holocene optimum of the mid-sub-Atlantic phase (1800 to 800 yr) was characterized by a larger amount of atmospheric precipitation, lower summer temperatures, and warmer winters. This optimum was followed by a period of lower humidity, when the climatic environment acquired its recent features.

Reviewer M.N. Alekseev.

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BRIEF COMMUNICATIONS

Findings of Small Shelly Fossils in Lower Paleozoic Deposits of the Talas Ala Tau (Northern Tien Shan)

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Abstract – Recent paleontological data on the ancient deposits of the Talas Ala Tau contradicts accepted viewpoints on their stratigraphic subdivision. By using chemical preparation in a detailed stratigraphic study, small shelly fossils (chiolites, hyiolitelmints, and sponges) as well as algae were obtained from the flysch sequences (the Chatkaragai Subgroup) of the Karagai Group of the northern slopes of the Talas Ala Tau. According to these findings, the age of carbonates enclosed in the Urmaral Formation of the Chatkaragai Subgroup is considered to be the Early Paleozoic (the undivided Cambrian–Ordovician). Earlier, these sediments were attributed to the Upper Riphean.

The application of chemical methods for rock disintegration, along with the use of small shelly fossils, add more detail to the paleontological characteristics of ancient "barren" deposits of the Tien Shan.

Biostratigraphic studies carried out from 1960 to 1970 resulted in findings of Early Cambrian fossils (Keller and Pokrovskaya, 1965; Missarzhevskii and Mambetov, 1981) in phosphoritic layers of the Malyi Karatau, the northwestern spurs of the northern Tien Shan. Deposits underlying beds with Early Cambrian organic remains have been referred to the Precambrian (Korolev, 1962; Keller *et al.*, 1965; Korolev and Maksumova, 1964, 1968). Isotopic data used for dating the Precambrian sequences of stromatolites and oncolites provided grounds for stratigraphic and correlation schemes of the Tien Shan, where the Riphean was divided into three units and the Vendian, into two units (Korolev *et al.*, 1974).

Recently, however, new data was published on fauna occurrences in some sequences assigned earlier to the Precambrian (Mitrofanov *et al.*, 1983; Maksumova *et al.*, 1987).

New data on the stratigraphy of some "Precambrian" sequences, formational analysis together with facies and genetic inferences, and reinterpretation of the contact relations between the Precambrian and Paleozoic sequences resulted in the concept of a complex imbricated nape structure of the Tien Shan (Maksumova *et al.*, 1988; Khudolei and Semiletkin, 1992; Alekseev, 1993). One of these complex structural zones is the Talas–Karatau region, where ancient sequences have been accepted as the Upper Precambrian and Lower Paleozoic reference sections of the northern Tien Shan.

Two types of Precambrian sections were distinguished in the Talas Ridge: the reduced Kurgan type section, and the complete Kumyshtag type section (Korolev and Maksumova, 1964). The Beshtash–Terek fault separates their distribution areas (Fig. 1). The upper parts of both sections are represented by very similar complexes of the Malyi Karoi, Dzhemtyn, and Tamda groups, with definite correlation. However, correlation of the underlying sequences is doubtful. In the reduced section, the lower part of these underlying sequences is composed of massive quartz-feldspar and quartz sandstones assigned to the Sarydzhon Formation (700 m) and by light rose limestones and dolomitic limestones of the Chatkaragai Formation (200 - 300 m). The upper part of the section comprises red and green siltstones, mudstones, and thin-platy limestones of the Kyzylbel' (50 - 200 m) Formation (Dodonova, 1957, 1960). In the limestones of the Chatkaragai Formation, Dyadyuchenko, Korolev, and Kiselev revealed stromatilites, attributed to the Upper Riphean Min'yar complex by Krylov (1967).

Areas with a similar sequence of strata but of a different type of formation were attributed to the complete Kumyshtag type section. A thick sequence of carbonate-terrigenous flysch (with terrigenous graywacke in the upper part) was correlated with the sandstone sequence of the reduced sections, and Dodonova assigned it to the Sarydzhon Formation. Its total thickness is 3000 m. Later, Maksumova subdivided this unit into the Tagyrtaus and Sarydzhon formations (Korolev and Maksumova, 1964), whereas other geologists (Bekker *et al.*, 1988) distinguished there the Chondzhol, Tagyrtaus, and Chidygolot formations. Dodonova attributed the thick (up to 3000 m) sequence of the carbonate flysch to the Chatkaragai Formation and correlated it with thin beds of massive stromatolitic limestones of the reduced Kurgan type section.

The variegated lower marine molassa (up to 2000 m) crowning the terrigenous and carbonate flysch was correlated by Dodonova with the Kyzylbel' Formation. Later, Maksumova (1980) subdivided this interval of

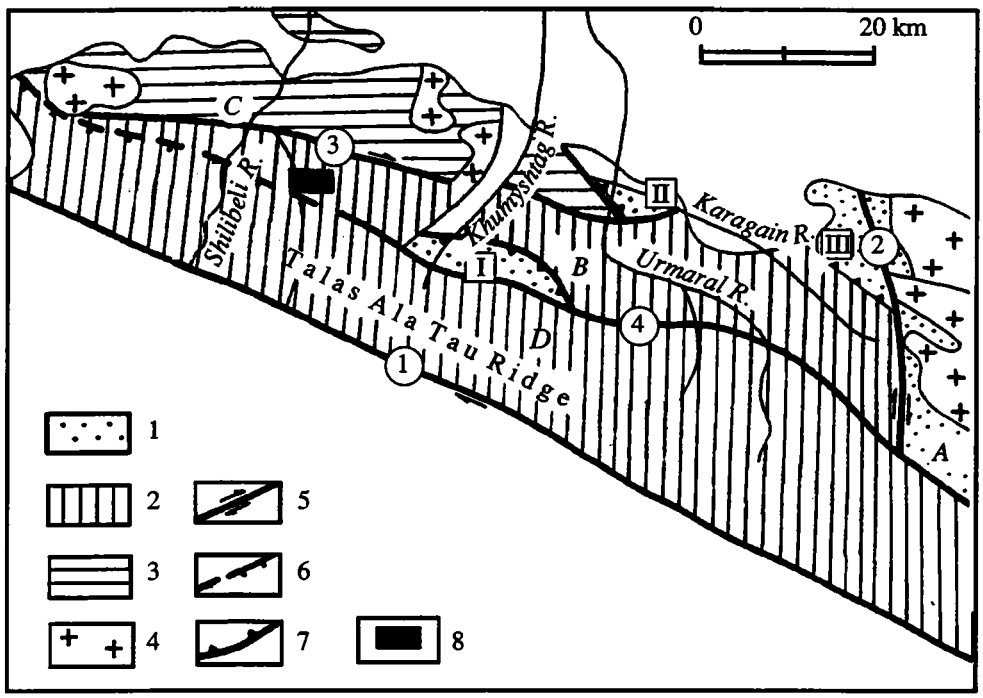

Fig. 1. Tectonic scheme of the Talas Alatau.

1 – distribution area of the "reduced" Kurgan type section of the Upper Precambrian; 2 – distribution area of the complete Kumyshtag type section of the thick terrigenous-carbonate and carbonate flysch; 3 – terrigenous flysch; 4 – granitoid intrusions; 5 – strike slip faults, 6 – thrusts, 7 – allochthons; 8 – the section site.

Faults: 1 – Talas-Fergana; 2 – Beshtash-Terek; 3 – Dzholsai; 4 – Uzunakhmat-Kumyshtag.

Blocks: *A* – Beshtash-Kolba; *B* – Kumyshtag; *C* – Tagyrtaus; *D* – Uzunakhmat.

Allochthons: I – Kumyshtag; II – Mailibulak; III – Karagain.

the section into the Chatkaragai and Sagyzgan formations (representing the undivided Chatkaragai Formation after Dodonova), and, because of its formation different type, the upper variegated part of the section was renamed as the Postunbulak Formation. The stratigraphic position of the latter was defined as higher than that of the Kyzulbel' Formation of the reduced Kurgan type section. Accordingly, the Kyzylbel' Formation was correlated with the Sagyzgan Formation.

During past years, mapping geologists denied the names Sarydzhon and Chatkaragai for the formation of this region and subdivided the thick sequence of carbonate flysch into the Birbulak, Urmaral, and Chokutash (= Sagyzgan) formations, though for subgroups of the Karagain Group they left the names given above.

The thick flysch sequence of the Kumyshtag type section lacked paleontological characteristics. Only data on nonrepresentative Upper Ripheran oncolites in limestones of the Urmaral Formation were available. The Late Riphean age of the flysch sequence was determined by its stratigraphic position below the Malyi Karoi Group of Vendian age (Korolev and Maksumova, 1968), and by correlation with the reduced Kurgan type section, characterized by Upper Riphean stromatolites. Later, in 1979 - 1980, thin-sections (collection of Maksumova and Malyuzhinets, Moscow State University) from limestones of the Urmaral Formation exposed along the Taity River, the right tributary of Kara-Buur River, were studied by Kringer (the Central Complex Geological Expedition, Karaganda), who found shell fragments of brachiopods and trilobites and abundant algae testifying to the Paleozoic age of enclosing rocks.

In 1985, Kras'kov described problematic remains of *Caraburina*, which were extracted by chemical preparation from samples collected in the above locality. He noted their similarity with Upper Riphean fossils from several regions of the world. However, the controversial interpretation of the *Caraburina* remains should be noted. Therefore, in the monograph *Precambrian Microfossils of the USSR* (1989) these remains are described as microfossils of faunal origin, but their similarity with lower plants and algae is also pointed out.

In 1986, Abduazimova (OME SAIGIMS, Tashkent) found early Paleozoic (Ordovician) Chitinozoa in thin-sections characterizing the entire "Riphean" section of the Talas Ridge. In 1985, Il'chenko presented an extended list of phyto-microfossils from the same samples, which were used to recognize all the Riphean units.

As evident from the aforesaid, the data on paleontological characteristics of the ancient sequences of the Talas Ala Tau are inconsistent and can not be used for the unambiguous age determination of these deposits.

During past years, we again tried to reveal fossils in rocks of thick flysch sequences. Naturally, the section of the Urmaral (or lower parts of Chatkaragai) Formation in the lower course of the Taity Stream (the right tributary of the Kara-Buura River, in its middle course) was chosen as an object for thorough study, because earlier, Kras'kov (1985) found microfossil remains and fragments of shell fauna in this location. The section described below is located in the Taity Stream mouth (Fig. 2), the right tributary of the Kara-Buura River (the left tributary of the Talas River) in northern spurs of the Talas Ala Tau.

A section of the lower carbonate subgroup of the Urmaral Formation was compiled by studying outcrops on both sides of the stream, 200 m higher than its mouth (Fig. 3). It consists of the following members:

1. Limestones, thick to moderately bedded with some signs of wavy bedding, dark grey, pelitomorphic, and fine-grained. Interlayers of thin-platy varieties and shales are available. Algae of the genera *Girvanella* and *Epiphyton* (determination by Luchinina, IG i GSO RAN) are encountered at the layer base. The thickness is 24.5 - 25 m.

2. Angular conglomerates, gravelstones, and sandstones, representing debris deposits, overlap the underlying rocks with uneven contact. The clastic material is represented by elongated tabular limestone fragments in a clay-carbonate matrix of a contact cementation type. There are interlayers (up to 0.5 m thick) of fine-grained limestones, mudstones, and shales. The cementing mass contains no quartz and feldspar grains of sandy or silty size. Scarce remains of *Hyolithelus* sp. are present at the base of the member, in the fine-grained limestone layer. Algae of the *Girvanella* and *Epiphyton* genera are abundant in the middle and upper parts. The thickness is 21 m.

3. Limestones, dark-grey, fine-grained, with moderate or weak wavy-bedding. The lower part of the member comprises algae of *Girvanella* genus of the Cambrian affinity (not determinable at the species level because of high recrystallization). Remains of sponges of the class Hexactinellida were extracted from this interval by chemical preparation. The thickness is 11 m.

4. A bed of angular conglomerate of carbonate composition rests with traces of weak erosion on the limestones. Upward, it grades into thin-platy mudstone and greenish-grey shale. Higher, there are dark-grey and fine-grained limestones with irregular bedding. The thickness is 16 m.

5. Brown-grey, sandy-silty clastic limestones (calcarenites) occur above the uneven erosion surface. They include rare, poorly rounded pebbles and gravel-size fragments of dark-grey limestones. Upward, the rocks grade into fine-grained carbonate sandstones

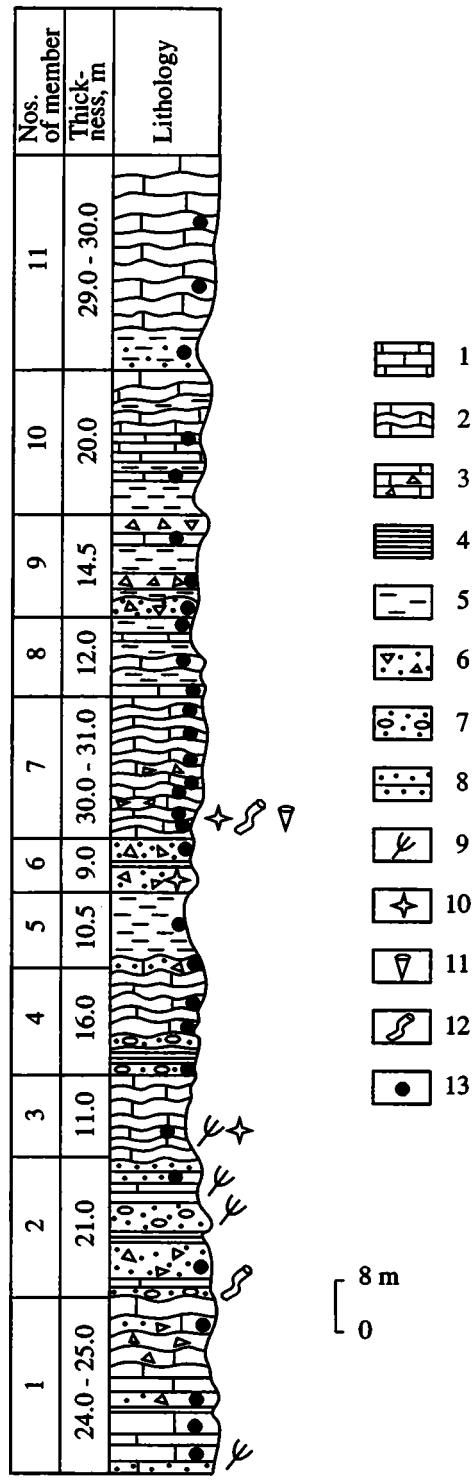

Fig. 2. Geological section through the Umaral Formation of the Chatkaragai Subgroup in the Kara-Buura River locality (the Taity River mouth).

1 – platy limestone; 2 – irregularly laminated limestone; 3 – clastic brecciated limestone; 4 – shale; 5 – mudstone; 6 – angular conglomerate; 8 – sandstone.

Fossils: 9 – algae; 10 – sponge spicules; 11 – chiolites; 12 – hyiolitelmints; 13 – intervals sampled for dissolution analysis.

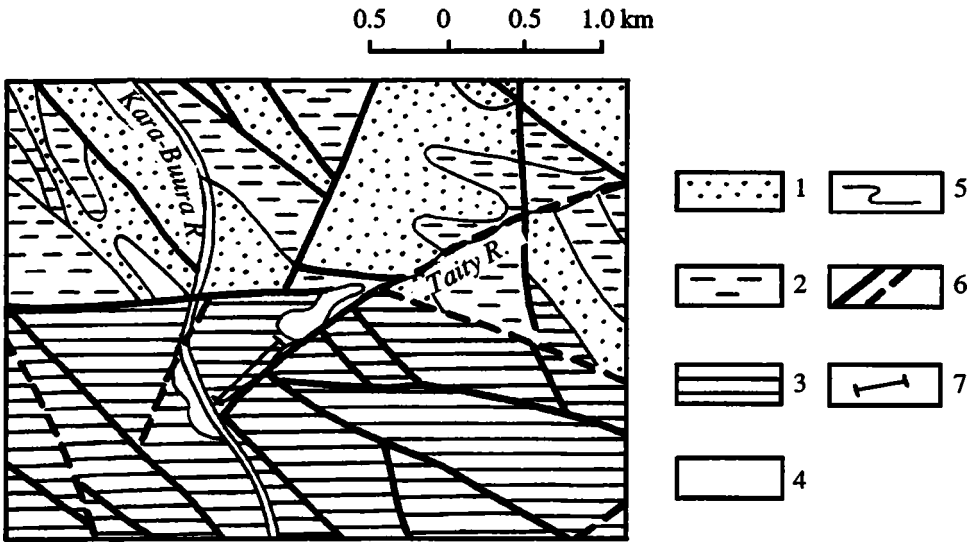

Fig. 3. Geological map of the middle course of the Kara-Buura River. Scale 1 : 50 000 (compiled by Konyukhov, 1992).
1 – Tagyrtaus Formation; 2 – Chidygolot Formation; 3 – Urmaral Formation; 4 – Quaternary deposits; 5 – stratigraphic contacts; 6 – tectonic contacts; 7 – geological cross section.

with an admixture of quartz-feldspar clastic material. The sandstones are overlain by carbonate mudstones with abundant biotite flakes on the bedding planes. The thickness is 10.5 m.

6. Angular calcareous conglomerates, rusty-brown with obscure bedding. Their fragments are represented by rod-tabular limestones in a sandy calcareous cement. Higher in the section, they are generally moderately bedded and include mudstone interlayers. The thickness is 8.5 - 9 m.

7. Limestones, irregularly bedded, fine-grained. They reveal intervals subjected to strong silicification and include abundant interlayers (up to 0.7 m thick) of calcarenite and clastic oncolite limestones. In the lower part of the member, there are individual remains of poorly preserved chiolites defined as Allacthetidae gen. et sp. indet., worm tubes of *Hyolithellus* sp. indet., and siliceous sponges of the class Hexactinellida. The thickness is 30 - 31 m.

8. Rhythmic alternation of platy mudstones and thick-platy limestones with wavy bedding. The upper part of the member is represented by mudstones with interlayers of fine-grained limestones. The thickness is 11.2 - 12 m.

9. Calcareous sandstones and thin-platy siltstones occur above the uneven contact and upward grade into clastic, brecciated dark brown limestones. The clastic material is represented by rodlike and tabular limestones and mudstones. The cement is fine-grained, calcareous, with rare gravel-sandy clasts. Higher in the section, they grade into highly foliated mudstones and then into conglomerates (up to 4 m thick) with rare angular boulders. The latter are overlain by thin-platy mudstones alternating with fine-grained limestones. The top of the member consists of clastic brecciated limestones with rod fragments. The thickness is 14.2 - 14.5 m.

10. Greenish-grey mudstone with occasional thin interlayers of limestones, sandstones, and siltstones, grading higher into fine-grained brownish-grey limestones with regular bedding. The rocks include interlayers of their clastic brecciated varieties. The uppermost part consists of alternating mudstones and limestones. The thickness is 19.7 - 20 m.

11. Siltstones and mudstones, thin-bedded, with rare interlayers of limestones. The member is crowned with monotonous layered brownish-grey limestones. The thickness is 29.0 - 30 m.

After dissolution of many samples (more than 100) collected from the entire section of the Urmaral Formation described above, we revealed small shelly fossils from several stratigraphic levels (Fig. 3). They are represented by rare, poorly preserved remains of animal and plant origin. These are chiolites of the family Allathecidae (layers 7 and 6; Fig. 3), worm tubes of the genus *Hiolithellus* Bill., (layers 2 and 7), and sponges of the class Hexactinellida (layers 3 and 7). In addition, fossil algae of the *Girvanella* and *Epiphyton* genera are abundant in the lower part of the section (layers 1, 2, and 3; Fig. 3). The latter were detected in petrographic thin-sections and obtained with chemical disintegration of rocks.

According to Luchinina (IG i GSO RAN), who reviewed our collection, algae of the *Girvanella* genus are represented by new forms not yet described in the literature, similar to those from the Cambrian deposits of the Altai–Sayan region and Siberia. Due to poor preservation (high recrystallization), algae from the group of epiphytes are not determinable at the species level.

To confirm the data obtained, the same part of the section along the left side of the Taity Stream was studied in detail. As it does not differ from the above section, there is no need to describe it here. We should note only that small shelly fossils were also revealed in this section, starting from its base. Here, numerous sponges from the class Hexactinellida and individual hyiolitelmints of the genus *Chiolitellus* are distinguished. Small shelly fossils bear clear traces of redeposition (they are represented by fragments only). This explains the scarcity of findings.

According to new findings of chiolites, hyiolitelmints, sponges, and algae, the age of the lower carbonate subformation of the *Chatkaragai* Formation (the Urmaral Formation, according to Bekker *et al.*), which we described above, can confidently be defined as the early Paleozoic instead of the Late Riphean, which was suggested earlier.

According to the vertical distribution in the section and on the basis of available data (Missarzhevskii, 1989; Bengston *et al.*, 1990), the age of small shelly fossils (chiolites of the family Allathecidae, hyiolitelmints of the genus *Hyolithellus*, sponges of the class Hexactinellida) can certainly be attributed to the undivided Cambrian–Ordovician. The scarcity of fossils in the rocks seems to be a result of their poor preservation in turbidite and debris flow deposits. Such features are typical of the younger flysch formations as well.

The first findings of fossils from "*barren*" sequences of the northern Tien Shan, assigned earlier to the Precambrian, suggest the desirability of a detailed biostratigraphic study on small shelly fossils and the redefinition of their age.

Reviewer A.Yu. Rozanov

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Paleocene Nummulites from the Crimea, Caucasus, and Middle Asia

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Our work was initiated by the late academician Menner, who had always been interested in the Paleocene nummulites, and especially occurrence in the CIS.

It was long thought that in the southern provinces of the former USSR nummulites appeared only since the earliest Eocene. However, they were found in older Paleocene deposits due to detailed studies on these larger foraminifera in many localities of the indicated regions. These studies were initiated at the end of the 1940s, and now nummulites are recognized in all stratigraphic units of the Paleocene deposits of the southern CIS, namely in their Danian, Montian, and Thanetian stages (Fig. 1).

In 1989, Ashurov detected Danian (lower Paleocene) nummulites in the following two sections of the Bakhchisarai area, southwestern Crimea: (1) along the Bel'bek River, near the village of Tankovoe, and (2) on Mt. Besh-Kosh, near the village of Skalistoe. In both sections, nummulites occur in hard bryozoan-crinoidal limestones of the upper Danian, bearing inarticulate brachiopods *Ancistocrania tuberculata* (Nills.) and *A. ignabergensis* (Retz.) along with abundant columnals of Crinoidea species *Bourgueticrinus* sp. The rocks yielded rare and very small nummulite tests, including those of three new species: *Nummulites danicus* Ashurov and Nemkov, *N. belbekensis* Ashurov and Nemkov, and *N.*(?) *unicalis* Ashurov and Nemkov (Ashurov and Nemkov, 1991). No data on nummulite occurrences in the Danian deposits had been reported previously either in this country or abroad.

Occasional specimens of the Montian nummulites were discovered in the northern Caucasus, Transcaucasia, and Central Asia. In the Kuban' River basin, northern Caucasus, Bugrova (1984) recognized primitive forms of *Nummulites* aff. *deserti* de la Harpe. Their rare tests were found at the top of the El'burgan Formation in association with smaller foraminifera of the *Globorotalia conicotruncata* Zone and within the *Acarinina djanensis* Zone of the middle part of the Goryachii Klyuch Formation (Bugrova, 1984). Revising this collection in the Chernyshov Museum of the All-Russian Geological Institute, we realized that discovered tests belong to nummulites of an unclear specific type. Earlier, in 1964, Kacharava and Kacharava described *Nummulites fraasi* de la Harpe, *N. solitarius* de la Harpe, *Discocyclina seunesi* Douvill, *D. douvillei* (Schlumb.), and *Operculina* sp. from clay beds of the Algeti River, where these sediments overlie the variegated Danian deposits. Using this data, Mrevlishvili (1978) assigned the rocks to the lower Paleocene. Some nummulite species were also found in the Akdzhar Horizon (lower Paleocene) deposits of the Tadzhik depression (Mrevlishvili, 1978) and were determined by Ashurov to be *Nummulites fraasi* de la Harpe, *N. deserti* de la Harpe, and *N. solitarius* de la Harpe (Ashurov and Nemkov, 1978).

Nummulites are widely distributed in the upper Paleocene deposits. Their richest assemblage is known from the Tadzhik depression. In the Tabakchi Horizon of this area, Ashurov encountered three *Operculina* forms and 25 species and subspecies of Nummulitidae, including *Nummulites fraasi* de la Harpe, *N. fraasi densispira* de la Harpe, *N. deserti* de la Harpe, *N. solitarius* de la Harpe, *N. silvanus* Schaub, *N. praecursor alpinus* Schaub, *N. soerenbergensis* Schaub, *N. bolcensis* Mun.-Chalm., *N. spileccensis* Mun.-Chalm., *N. panteleevi* Nemkov, *N. subplanulatus* Hant. et Mad., and others (Ashurov and Nemkov, 1978).

The Transcaucasia is another region of abundant upper Paleocene nummulites. Mrevlishvili (1978) described *Nummulites fraasi* de la Harpe, *N. solitarius* de la Harpe, *N. silvanus* Schaub, *N. soerenbergensis* Schaub, *N. praecursor* de la Harpe, *N. pernotus* Schaub, and *Discocyclina seunesi* Douvill from Georgia. Most of these species occur in the Algeti River basin.

Species of upper Paleocene nummulites are recognized also in Azerbaidzhan. Here, Mamedov (1967) classified *Nummulites fraasi* de la Harpe, *N. deserti* de la Harpe., *N. subplanulatus* Hant. et Madar., *N. bolcensis* Mun.-Chalm., *N. exilis* Douvill, *N. nitidus* de la Harpe, *N. praeexilis* Mamedov, and *N. djulfensis* Mam..

Single nummulite species were described from the upper Paleocene deposits of the Ukrainian Carpathians, Crimea, Armenia, Mangyshlak, central Kyzyl Kum, eastern Kopet-Dag, Gyaur-Dag, and Badkhyz (Golev, 1980; Grigoryan, 1986; Nemkov, 1967).

All the data suggest that nummulites have a nonuniform distribution in the Paleocene deposits of the entire

area under consideration. They are most abundant in Central Asia, namely in the Tadzhik depression, where their species sometimes occur as rock-forming organisms. This implies that during the Paleocene the Tadzhik depression served as a marine basin that favored a nummulite habitat. Simultaneously, this contradicts some theories that there are no nummulites occurring east of Badkhyz.

The abundant nummulites from the Paleocene deposits of the Tadzhik depression are of great scientific interest, because their Paleocene assemblages of similar diversification have not been recognized elsewhere.

It is essential that in the southern CIS, in western and southern parts of the Mediterranean geosynclinal belt for example, nummulite assemblages include the well-known and typical Paleocene species, such as *Nummulites fraasi* de la Harpe, *N. fraasi densispira* de la Harpe, *N. deserti* de la Harpe, and *N. solitarius* de la Harpe. These forms are vital in solving problems of correlation and stratigraphy of the Paleocene deposits.

In the Paleocene deposits, the indicated forms occur along with subtransitive species (Golev, 1980), which were distributed all over the Mediterranean area mainly in the early Eocene. According to Schaub (1981), these and a number of other forms are typical of the Ilerdian Stage (upper Paleocene–lower Eocene) and presumably have appeared only since this time.

In fact, analyzing the vertical distribution of some nummulite groups appearing at the beginning of the Ilerdian time, one may refer the lower part of the Ilerdian to the uppermost Paleocene. In such a case, the Paleocene–Eocene boundary should be placed between the Ilerdian horizons "moyen-1" and "moyen-2" of Schaub's subdivisions. Precisely from this moment, the main groups of the early Eocene nummulites began to evolve (Schaub, 1981; pp. 28 - 30, Figs. 23 - 25) and, according to the calcareous nannoplankton, here we have the boundary between the *Discoaster multiradiatus* and *D. binodosus* zones (Andreeva-Grigorovich, 1980).

Studies on nummulites in the southern CIS and some localities of the eastern Mediterranean show that these larger foraminifera of the Paleogene existed even before the Paleocene in the late Maestrichtian (Ashurov *et al.*, 1987; Grigoryan, 1986; Dzhalilov *et al.*, 1988).

During the last 25 years, nummulites have been used for Paleocene stratigraphic zonation. According to resolutions of the Interdepartmental Stratigraphic Committee on the Paleogene System (*Resheniya ...*, 1967; pp. 51 - 54), the *Nummulites fraasi* Zone is accepted as the upper part of the Paleocene. This zone is recognized in the Crimea, Armenia, and Mangyshlak. In the Paleocene deposits of Georgia, Mrevlishvili (1978) detected two zonal units: the lower *Nummulites fraasi* Zone and the *Nummulites silvanus* Zone above it.

We (Ashurov and Nemkov, 1978) subdivided the rich nummulite assemblage of the Tadzhik depression into two units as well: the *Nummulites fraasi* Zone of the lower Paleocene, bearing a group of *N. fraasi, N. fraasi densispira, N. deserti*, and *N. solitarius*; and the *Nummulites silvanus–N. fraasi* Zone of the upper Paleocene, which includes such species as *N. silvanus, N. soerenbergensis, N. pernotus, N. bolcensis, N. subplanulatus, N. spileccensis, N. praecursor alpinus, N. solitarius, N. deserti*, and *N. fraasi*.

In the upper Paleocene deposits of Azerbaidzhan, Mamedov also recognized the *Nummulites silvanus–N. fraasi* Zone with abundant nummulite species. In the Ukrainian Carpathians, Crimea, Armenia, Mangyshlak, and another regions, nummulites of the lower Pale-

Regions	Tadzhik depression			Central Kyzyl Kum	Badkhyz	Gyaur-Dag	Eastern Kopet-Dag	Mangyshlak	Kuban'			Azerbaidzhan	Armenia	Georgia		Crimea		Ukrainian Carpathians
Stratigraphic units / Nummulite species	Akdzhar Horizon	Tabakchi Horizon	Aruk Tau Horizon	Upper Paleocene				Suluk Apa Formation	El'burgan Formation		Goryachii Klyuch Formation	Upper Paleocene		Lower Paleocene	Upper Paleocene	Danian Stage	Upper Paleocene	
Nummulites fraasi	+	+	+	+	+	+		+				+	+	+	+		+	+
N. fraasi densispira		+											+					
N. solitarius	+	+	+				+						+					
N. deserti	+	+	+						+	+		+	+					+
N. silvanus		+													+			
N. panteleevi		+																
N. bolcensis		+										+						
N. pernotus		+													+			
N. spileccensis		+	+									+						
N. spileccensis regulata		+																
N. praecursor alpinus		+																
N. praecursor															+			
N. praeexilis		+										+						
N. exilis		+										+	+					
N. soerenbergensis		+													+			
N. globulus		+																
N. subplanulatus		+										+						
N. nitidus												+						
N. aff. mouratovi		+																
N. akkuurdanensis		+																
N. istimtauensis		+																
N. tadjikistanensis		+																
N. barkhatovae		+																
N. salibaevi		+																
N. (?) guliensis		+																
N. (?) babkovi		+																
N. djulfensis												+						
N. danicus																+		
N. belbekensis																+		
N. (?) unicalis																+		
Operculina parva		+																
O. ex. gr. heberti		+											+					
O. ex. gr. canalifera		+																
Discocyclina seunesi													+	+			+	
D. douvillei														+				

Fig. 1. Distribution of nummulites in the Paleocene deposits of the southern CIS.

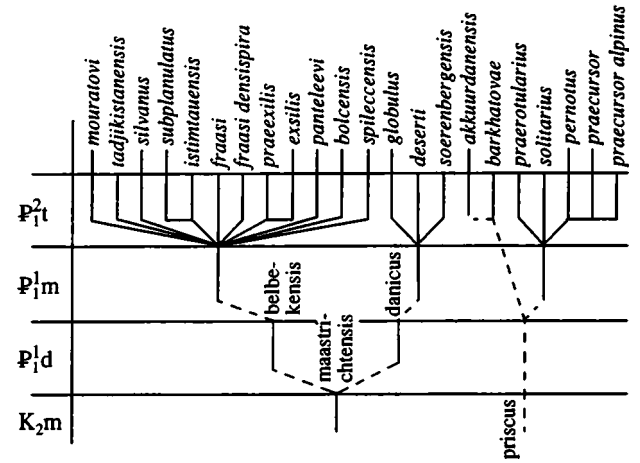

Fig. 2. Provisional phylogenetic scheme for the Late Cretaceous and Early Paleogene nummulites occurring in the southern CIS.

ocene deposits are still unknown, and this complicates correlation of corresponding rocks.

Summarizing the presented data, we may conclude that in a vast area of the southern CIS there are many occurrences of Paleocene nummulites. A detailed study on these larger foraminifera will be useful for correlating sections of these countries and East Mediterranean regions.

In closing, we present a phylogenetic scheme for the Paleocene nummulites (Fig. 2). As it is based upon our own data, it is provisional and requires additional verifications.

Reviewer A.L. Yanshin

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BRIEF COMMUNICATION

Biostratigraphy of the Khazarian Beds in the Seroglazovka Reference Section, Northern Caspian Region

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Marine Khazarian sediments of the Caspian region are usually subdivided into upper and lower beds. Pravoslavlev (1926) was the first to show this. Zhukov (1935) and Fedorov (1957) confirmed his supposition. However, sections with joint occurrences of fossiliferous upper and lower Khazarian beds are rare, and their biostratigraphly is poorly studied. One of the most complete sections of the Khazarian Beds in the northern part of the Caspian region is exposed in a scarp on the right bank of the Volga River at a distance of 1 km upstream from the Seroglazovka settlement. Here, for a distance of two kilometers, (Fig. 1) a multilayered polyfacial series of marine sediments with various mollusk and ostracode fossils is exposed.

The following members are distinguished in the section (from the top downward):

1. Dusting yellow-gray, fine-grained sand, 0.5 - 7.0 m thick. The lower boundary is sharp and unconformable. Four fossil cultural layers are traced in the thickest, western part of the exposure. They are represented by coal remnants, broken ceramics, and other household remnants. Radiocarbon age of the coal is 505 ± 68 (MGU–1755) years.

2. Dusting dense sand, brownish and yellow-brown, thin-laminated, cross-bedded, with a slightly humified soil horizon at the bottom. In the most complete parts of the layer, there are two interlayers separated by weathering traces: the upper one colored light yellow-brown, and the lower one of a more dark brown color. Molluscan shells of *Didacna protracta, Dreissena rostriformis*, and rare *Sphaerium* forms were found here together with the brackish- and freshwater ostracodes *Cyprideis torosa, Caspiolla gracilis, Aurilla azerbaidjanica, Leptocythere bacuana*, and *Iliocupris bradyi*. The member is 5 m thick. The lower boundary is sharp, sometimes marked by deep wedges cutting the underlying layer.

3. Brownish-gray loamy sand with brown interlayers and traces of soil formation at the bottom. The rocks bear rare detritus of molluscan shells. The thickness of the member does not exceed 1.5 m.

4. The member has a complicated structure, especially in the eastern part of the exposure (Fig. 2), where it includes several lithofacies interlayers. At the bottom of the layer (Bed a) there are fine gray sands, well-sorted, with thin cross-bedding and abundant shells of *Didacna surachanica, D. nalivkini, D.* cf. *pontocaspia*, and more rare *D. schuraosenica, D.* aff. *pallasi, Dreissena caspia, Dr. rostriformis, Micromelania caspia, Monomania caspia*, and *D. zhukovi*. These sands unconformably overlie underlying deposits. Upward, they are replaced by greenish-gray nonstratified sands, slightly ferruginate (Bed b) and passing upward into dove-colored loamy sand slightly affected by processes of hydromorphic soil formation (Bed c). Along the strike, the last bed is replaced by alternating brown clays and gray sands (Bed d). Beds a - c form a residual form of buried relief unconformably overlain by gray cross-bedded sands (Bed e). The latter contain abundant molluscan shells of *Dreissena rostriformis, Monodacna caspia, Didacna* ex gr. *trigonoides, D.* cf. *pontocaspia*, more rarely *Didacna surachanica, D. nalivkini, D. subcrassa, D. pallasi, Dreissena caspia, Dr. polymorpha, Hyphania plicata, Micromelania caspia, Corbicula fluminalis*, and *Sphaerium*, and scarce *Didacna schuraosenica, D. zhukovi, Theodoxus*, and *Unio*. Upward, these sands are replaced by light-yellow horizontally laminated sands. They include a lentil of light-brown and dirty-yellow sand (Bed g) with loamy sand interbeds, cross and lenslike-bedded, and basal interbeds of clayey rounded pellets and calcareous concretions. The rocks contain abundant shells of the fresh- and brackish-water mollusks *Unio pictorum, Corbicula fluminalis, Sphaerium rivicola, Dreissena polymorpha, Dr. celekenica, Didacna pallasi, D.* ex gr. *trigonoides*, and *Micromelania caspia* and rare *Didacna surachanica* and *D. schuraosenica*. Abundant fresh and brackish-water ostracodes of Member 4 include such forms as *Caspiolla gracilis, Iliocypris bradyi, Il. bella, Limnocythere braevis, Cypria candonaeformis, Bacunella dorsoarcuata, Paracyprideis naphtatacholana, P. enucleata, P. astrachanensis, Tyrrhenocythere pseudoconvexa, T. tenera, Aurilla azerbaidjanica, A. papillosa*, and *Leptocythere bacuana*. The member is 5 m thick.

5. Horizontal interbedding of brown and brownish-gray clays and loams with sands and gray silts, fine-grained and well-sorted. The upper interlayers are often deformed into folds to fill deep pot-holes. The upper boundary of the member is sharp with traces of considerable erosion. The lower one is lithologically distinct

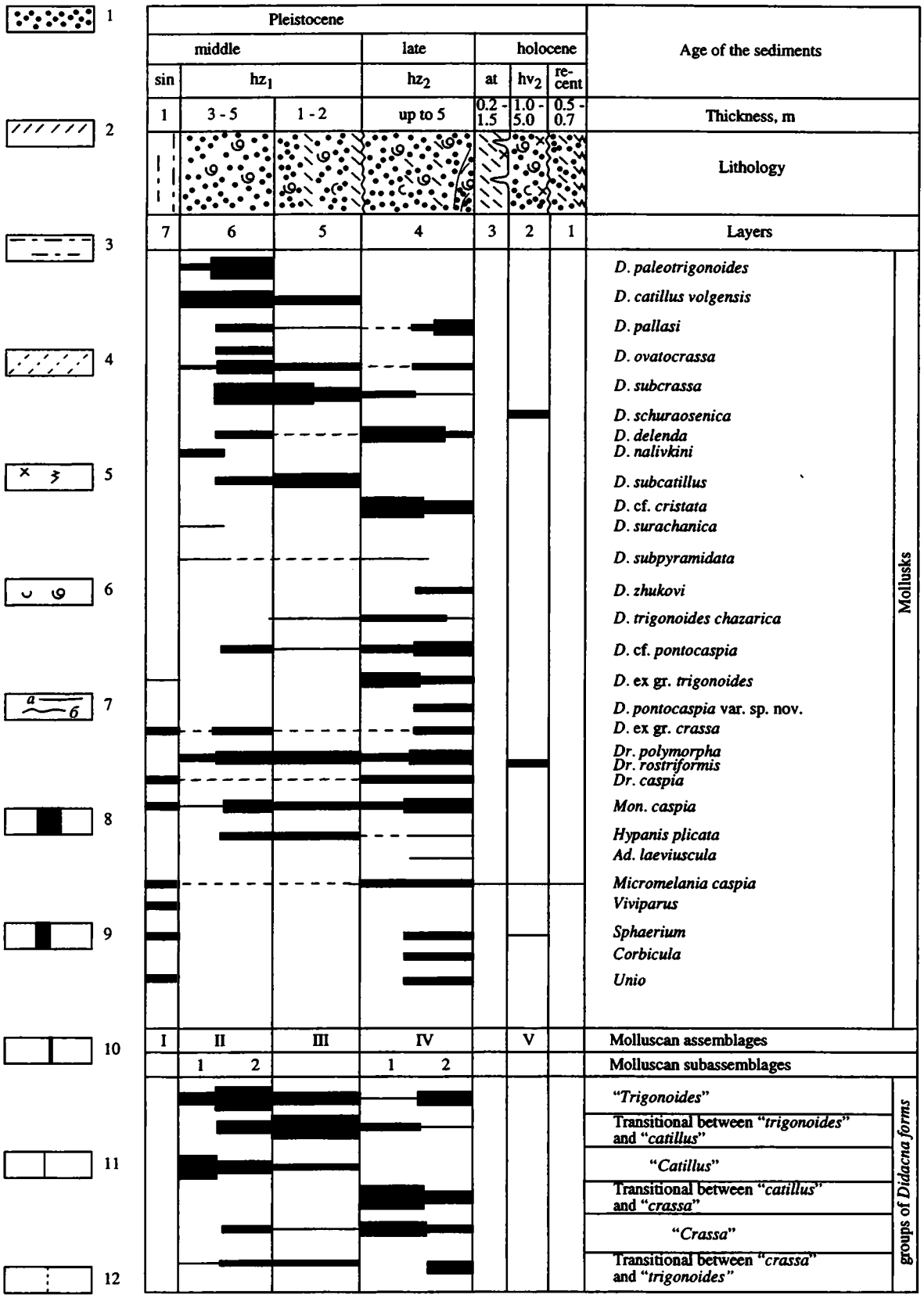

Fig. 1. Seroglazovka section.

1 – sand; 2 – loam; 3 – silt; 4 – loamy sand; 5 – fossil and recent soil; 6 – molluscan shells and shell detritus; 7 – boundary: a) conformable, b) unconformable; 8 – shells of the dominant species (group); 9 – abundant; 10 – rare; 11 – scarce; 12 – presumably existing.

and conformable. Shells of *Didacna schuraosenica*, *D.* cf. *cristata*, and *Dreissena rostriformis* are abundant, while those of *Didacna catillus volgensis*, *D. subcrassa*, *D.* cf. *pontocaspia*, *D.* ex gr. *trigonoides*, *D. pallasi*, *Hypanis*, and *Monodacna* are rather rare. Specimens of *Didacna schuraosenica* predominate in the lower part of the layer; upward they are accompanied by abundant *D.* cf. *cristata* and *Dreissena rostriformis*. Fossil ostracodes of the member include *Cyprideis torosa, Bacunella dorsoarcuata, Caspiolla gracilis, Cryptocyprideis bogatschovi, Paracyprideis naphtatscholana, P. enucleata, Tyrrehenocythere pseudoconvexa, Aurila azerbaidjanica, A. papillosa, Leptocythere bacuana, L. martha*, and *L. arevina*.

6. Yellowish-gray and light-gray fine sands, thin laminated, with brownish-gray interbeds of muddy sand, dense at the bottom and partly cemented by carbonates into platy sandstone with coquina interlayers and lenses. In the upper part of the member, the following fossils were determined: *Didacna paleotrigonoides* and *D. schuraosenica* along with abudant *D. catillus volgensis*, *D. subcrassa*, *D.* cf. *cristata*, *Monodacna*, *Hypanis*, *Micromelania caspia*, *Dreissena rostriformis*, common *Didacna ovatocrassa*, and rare *D. pallasi*, *D.* ex gr. *trigonoides*, *D. nalivkini*, *Dreissena polymorpha*, and *Theodoxus*. Its middle part is rich in *Didacna catillus volgensis* accompanied by common *D. paleotrigonoides*, *D. subcatillus*, and *Dreissena rostriformis*, and rare *Didacna subpyramidata*, *D. subcrassa*, *D. zhukovi*, and *Monodacna*. Among numerous brackish-water ostracodes of the member, *Cyprideis torosa*, *Caspiolla gracilis*, *Paracyprideis naphtatscholana*, *Tyrrhenocythere pseudoconvexa*, *Leptocythere bacuana*, *L. arevina*, *L. quinquetuberculata*, *L. pirsagatica*, and *L. cellula* occur. The sands of the member form the modern river beach. The lower boundary is lithologically sharp, partly deformed, with sandy pockets penetrating into the underlying sediments. The member is 3 - 4 m thick.

7. Dense, muddy, brown, and dark brown silt, sometimes sandy silt with the shells of *Didacna* cf. *pontocaspia*, *Dreissena polymorpha*, *Dr. caspia*, *Micromelania caspia*, *Unio*, *Viviparus*, and *Sphaerium*; species of *Didacna* are considerably stunted. The apparent thickness of the member is about 1 m.

According to lithological characteristics, position in the section, and fossils, the sediments of Member 1 are eolian deposits forming partly deflated small sandy ridges. Sediments of Member 2 represent deposits of the late Khvalinian sea basin. Underlying sediments of Member 3 are either the lower Khvalinian or the Atel' deposits. Taking into account the fact that in the Volga lower stream region the first deposits are usually represented by chocolate clays while the second show wedgelike deformations at the top (Vasil'ev, 1961), Member 3 corresponds most likely to the Atel' beds. The Blake episode of reversed polarity was determined in Member 4 (Eremin *et al.*, 1981). The composition of the layer suggests three stages in its formation, separated by deep erosional incisions. The first stage resulted in the accumulation of sediments composing the buried relief form. It started with the accumulation of marine sediments (Bed a), followed by lacustrine-liman (Bed b) and then lacustrine-bog sedimentation (Bed c). The second stage started after the erosional incision resulting in the formation of littoral-marine, deltaic, and shallow-marine deposits (beds d and e). A subsequent deep erosional incision was followed by the accumulation of alluvial and deltaic sediments in the erosion groove (Bed f). Molluscan assemblage from the Member 5 attributes it to the early Khazarian time. Popov (1983) correlated these deposits with the Girkan Beds, Sedaikin (1988) considered them to be of early Khazarian age, while Shkatova (1973) thought the silts of Bed 5 and the sands of Bed 4 to be the stratotype of the lower Khazarian marinine member at the Seroglazovka locality. According to fossil mollusks, the sands of Member 6 are referred to the marine lower Khazarian deposits. Fossil freshwater mollusks of the underlying silt layer suggest the latter to be terrestrial subaquatic sediments of Singilian age.

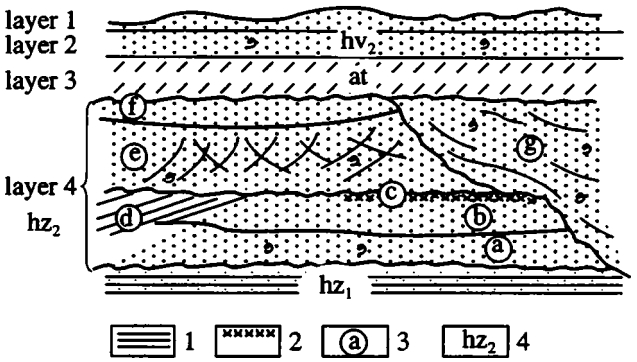

Fig. 2. Structure of Member 4 in the eastern part of the Seroglazovka section.

1 – clay; 2 – hydromorphic soil; 3 – numbers of lithological members; 4 – geological indexes: hz – Khazarian, at – Atelian, hv_2 – upper Khvalian deposits. Other symbols as described in Fig. 1, and in the text.

Malacofauna from members 6 - 4 is a typical Khazarian one (Fedorov, 1957) and includes numerous index-fossils of the early Khazarian (*Didacna paleotrigonoides*, *D. schuraosenica*, and *D. subpyramidata*) and late Khazarian (*D. surachanica*) epochs found *in situ*. The distribution of both index and diagnostic mollusk species within the section is uneven. The greatest differences are observed between mollusk compositions of members 6 - 5 and 4, correspondingly representing the early and late Khazarian faunas.

The early Khazarian fauna consists of assemblages II and III composed of 20 species. Mollusks of Member 6 comprise the assemblage II (*Didacna paleotrigonoides–D. schuraosenica–D. catillus volgensis*). In addition to these dominant species, it includes abundant

Didacna subcrassa, *D.* cf. *cristata*, *Monodacna caspia*, *Micromelania caspia*, and *Dreissena rostriformis* with common *Didacna subcatillus*, *D. subpyramidata*, *D. ovatocrassa*, *D. pallasi*, *D.* ex gr. *trigonoides*, *D. nalivkini*, *Hypanis plicatus*, and *Dreissena polymorpha*. It is interesting to note that *Didacna catillus volgensis* typical of the Baku deposits in the Volga lower stream region (Svitoch, 1966) is characteristic of the lower part of Member 6, while *D. paleotrigonoides* and *D. schuraosenica* predominate in its upper part. These changes in the mollusk compositions within Member 6 subdivide assemblage II into *Didacna catillus volgensis* and *Didacna schuraosenica–D. paleotrigonoides* subgroups.

The *Didacna schuraosenica* assemblage (III) featuring sediments of Member 5 includes abundant *D.* cf. *cristata* and *Dreissena rostriformis* with rare *Didacna catillus volgensis*, *D. subcrassa*, *D.* cf. *pontocaspia*, *D.* ex gr. *trigonoides*, *Monodacna caspia*, and *Hypanis plicata*. As a whole, mollusks from members 6 and 5 seem to represent different assemblages of the early Khazarian fauna.

The molluscan fauna of Member 4 is of the late Khazarian time and consists of 25 species. The late Khazarian index species *Didacna surachanica* together with *D. pallasi*, *D.* ex gr. *trigonoides*, *D.* cf. *pontocaspia*, *D. nalivkini*, *Monodacna caspia*, and *Dreissena rostriformis* predominate among them. There are also *Didacna subcrassa*, *D. trigonoides chazarica*, *D. schuraosenica*, *D. zhukovi*, *Dreissena caspia*, *Dr. polymorpha*, *Hypanis*, *Adacna*, *Micromelania*, *Sphaerium*, *Corbicula*, and *Unio* species in this assemblage. The abundance of *D. surachanica* shells sharply decreases in the upper part of Member 4, while the number of species preferring freshwater environments, or typical of them, increases. This subdivides the late Khazarian assemblage (IV) into the lower *Didacna surachanica* and upper *Didacna* ex gr. *trigonoides–Monodacna–Dreissena* subgroups. Changes in their compositions mark a transition from the brackish-water environment to a freshwater one of the liman-deltaic type.

Thus, the Khazarian malacofauna of the Seroglazovka section is represented by two early Khazarian (II and III) assemblages and one late Khazarian (IV) assemblage. *Didacna* species of "*catillus*" (lower subgroup) and "*trigonoides*" (upper subgroup) predominate within the lower assemblage (II), and transitional *Didacna* species (intermediate between "*trigonoides*" and "*catillus*" forms) are typical of the middle assemblage (III). In the upper assemblage (IV), mollusks of the "*crassa*" group and related taxa are replaced by abundant "*trigonoides*" forms upward in the section. The described distribution of the Khazarian fauna within the section is close to that of the marine terraces on the Dagestan coast of the Caspian Sea (Svitoch and Yanina, 1986). The only difference there is the appearance of the *Didacna crassa nalivkini* assemblage in the upper part of the lower Khazarian beds, which is absent in the Seroglazovka section. This assemblage is widely distributed in the northernmost sections of the northern Caspian region, far from the northern limit of the late Khazarian sea. Its absence in the Seroglazovka section is related to the erosion of the upper part of the lower Khazarian sediments by the late Khazarian transgressive basin.

Fossil ostracodes from the Khazarian deposits of the Seroglazovka section are represented by three assemblages. Most of them are typical of the Caspian marine Pleistocene, reflecting various ecological and sedimentation environments. Member 6 yielded the *Caspiolla gracilis–Leptocythere* assemblage, dominated by species of the last genus of the Bakunian - Khazarian time. The most diverse and abundant fauna of the shelly *Xestoleberis–Loxoconcha–Leptocythere* assemblage was found in Member 5. It probably corresponds to the transgression maximum and consists of species that appeared during the Khazarian time: *Loxoconcha unodensa*, *L. immodulata*, *L. abundata*, *L. lepida*, *L. lanta*, *Leptocythere bacinica*, *L. lunata*, *L. vermita*, and others. The brackish- to freshwater ostracods of the *Caspiolla gracilis–Limnocythere–Iliocypris* assemblage are characteristic of Member 4. They differ from the lower Khazarian fauna of members 5 and 6 by a sharply decreased occurrence of the *Leptocythere* and *Loxoconcha* taxa, and by an abundance of fresh-water shelly *Iliocypris bradyi*, *Il. bella*, *Limnocythere braevis*, and others.

Thus, we conclude with the following:

1. The study of the macro- and microfossils of the Seroglazovka section allows us to divide it into lower (Members 6 and 5) and upper (Member 4) Khazarian beds with characteristic assemblages of mollusks and ostracods. They are separated by a sharp lithological boundary and unconformity. The lower marine Khazarian deposits consist of two members displaying a successive accumulation from the beginning (Member 6) untill the maximum of the transgression (Member 5). The upper regressive part of the section was probably eroded. The upper Khazarian deposits are represented by polyfacial sediments ranging from marine to deltaic, and marked by numerous erosional surfaces.

2. We do not support the idea that silts of Member 5 can be referred to the Girkan Beds (Popov, 1983), because they are overlain by older deposits with *Didacna surachanica*. Their late Khazarian age (Shkatova, 1973) is also doubtful.

3. The Seroglazovka section of the Volga lower stream region is the richest in fossils among all sections of the northern Caspian region. This allows us to consider it as the reference section of the marine Khazarian deposits not only for this area, but for the entire Caspian region.

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BRIEF COMMUNICATION

Evolution of the biosphere

Diatoms from Deposits of the Sevsk Mammoth Burial Locality (Bryansk Region)

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The locality of the mammoth remains discovered in 1988 is situated in the southeastern part of the Bryansk district (the figure) in an interstream relic within the Sev River valley, where its brow corresponds in height to the 8-m-high terrace.

Being the largest among European mammoth burials (Lavrov and Mashchenko, 1991), this locality is of great importance. Four thousand bones of 30 - 35 individuals were collected here, whereas in the Condovery locality of Great Britain, there are remains of only 3 - 4 individuals. Compared to the Berelekh locality of Yakutia, where more than 8000 bones of 156 animals have been discovered, this occurrence is less rich, but more interesting as it has a higher abundance (60%) of young impubertal specimens including complete skeletons of mammoth babies from two months to five years in age. In addition, the Sevsk mammoths are the smallest among the known examples. Apparently, it was most recent mammoth population in Europe (13.950 ± 70 and 13.680 ± 60 years according to ^{14}C dating).

The unique features of the Sevsk locality prompted comprehensive investigations.

A bone-bearing lens was recovered during the mining of a sand quarry. The entire section has 5 members (from the top downward):

	Thickness, m
1. Grey sand, fine-grained, laminated, with interbeds of brown loam	up to 2.8 m
2. Dark brown loam with thin interbeds of sandy loams	0.15 - 0.7 m
3. Alternating light grey fine-grained sands, sandy loams, muddy sands, and loams. The member, rich in mammal bones, lies in a lenslike fashion over the uneven erosion surface of the underlying deposits	up to 0.7 m
4. Light grey sands, fine-grained, well-sorted, horizontally laminated, showing local cryogenic dislocations. The member thinswestward	up to 3.5 m
5. Grey sandy loams, horizontally laminated, with interlayers of light grey sand and dove-colored clay, grading downward into grey massive clay	1.6 m

Various methods of archaeology, radiocarbon dating, and palynology along with diatom and microtheriological analyses have been used to understand the climatic and paleogeographic environments at the time of the bone bed formation.

Eight samples characterizing all beds of the section have been selected to study the diatom fossils. The samples were boiled in Perhydrol to remove organic components and then enriched using a heavy liquid according to the standard technique (*Diatomovye...*, 1974). The studied samples yielded 58 species and subspecies

of 18 diatom genera from classes Centrophyceae and Pennatophyceae (the figure). All the forms are rare, i.e., only 1 - 10 valves occur per specimen. Nevertheless, when assemblages are representative, such an amount is admissible for analysis, because specific ecological niches of diatoms allow us to distinguish groups characteristic of certain biotopes, even if specimens are somewhat poor. Species of the *Eunotia*, *Pinnularia*, *Cymbella*, and *Gomphonema* genera are the most diverse (*Opredelitel'*..., 1951).

Diatoms can be divided into planktonic and benthic ecological groups. Only a few planktonic species (exactly 3, i.e., 5.2% of the total number) occur in the assemblages under consideration. These are *Aulacosira italica*, *Stephanodiscus astraea* v. *minutulus*, and *Tabellaria fenestrata*; these species occur throughout the sequence and are more common in its lower part.

Most forms are represented by benthic species divided into epibiontic and epiphytic forms. Epibiontic forms represent 43.9% of the total species number, and include 25 taxa of 9 genera. The *Pinnularia* genus has the most diverse species, which occur in every specimen. Species of the *Navicula*, *Calonies*, and *Nitzschia* genera were only identified in the lower part of the section, within the depth interval 5.25 - 5.80 m.

Epiphytes are also significant in the assemblage composition. They comprise 29 species and varieties of 6 genera and compose 50.9% of the assemblage. The epiphyte *Eunotia* is worth noting as it occurs throughout the sequence. It is somewhat more diverse in the depth interval 3.90 - 5.80 m., i.e., within the bone-bearing lens and underlying deposits.

Diatoms include species indicating different habitats. The studied assemblage is dominated by palustrine and paludo-lacustrine forms, such as *Tabellaria flocculosa*, *Fragilaria construens* v. *venter*, *Diploneis ovalis*, *Pinnularia subcapitata*, *P. viridis*, *Neidium bisulcatum*, *Cymbella aspera*, and many species of the *Eunotia* genus occurring in various combinations throughout the section. Only one rheophilic species of *Navicula cryptocephala*, characteristic of running waters, was identified in a specimen from a depth level of 5.25 m in the section's lower part. Two edaphophytes, *Pinnularia borealis* and *Hantzschia amphioxys*, were encountered in the depth interval 3.40 - 5.80 m.

The studied diatom flora is a freshwater oligohaline, i.e., it inhabits basins with a salinity of 0 - 5‰. It includes mainly indifferent species (80.8%; salinity range 0.2 - 0.3‰), but halophobous forms (salinity less than 0.2%) occur as well. These are *Tabellaria fenestrata*, *T. flocculosa*, *Diatoma anceps*, *Eunotia faba*, *E. monodon* et v. *major*, *E. pectinalis* v. *minor*, and *E. praerupta* et var. The assemblage is clearly barren of halophytes (salinity range 0.3 - 0.6‰).

With respect to pH, the studied diatoms can be divided into four groups: acidophytes inhabiting basins with an acid water reaction (28.7%); alkalibionts of basins with alkaline waters (6.7%); alkaliphytes occurring under pH = 7, i.e., under neutral water reaction but attaining optimum development in alkaline waters (24.4%); and indifferent forms populating both the acid and alkaline waters (40.2%). Acidophytes occur in a rather high proportion relative to other species and become the main elements of the assemblage in the depth interval of 3.90 - 5.80 m, within the bone-bearing lens and underlying deposits. Species of the *Eunotia* genus with *Tabellaria fenestrata*, *T. flocculosa*, and *Pinnularia major* forms are the main indicators of acid water conditions (*Diatomovye* ..., 1974).

According to geographic distribution, there are cosmopolitan species (39.6%), forms of Boreal provinces (39.6%), and representatives of North-Alpine zones (20.8%). These amounts of cryophilic elements are too high for the temperate zone, and their percentage decreases upward in the section.

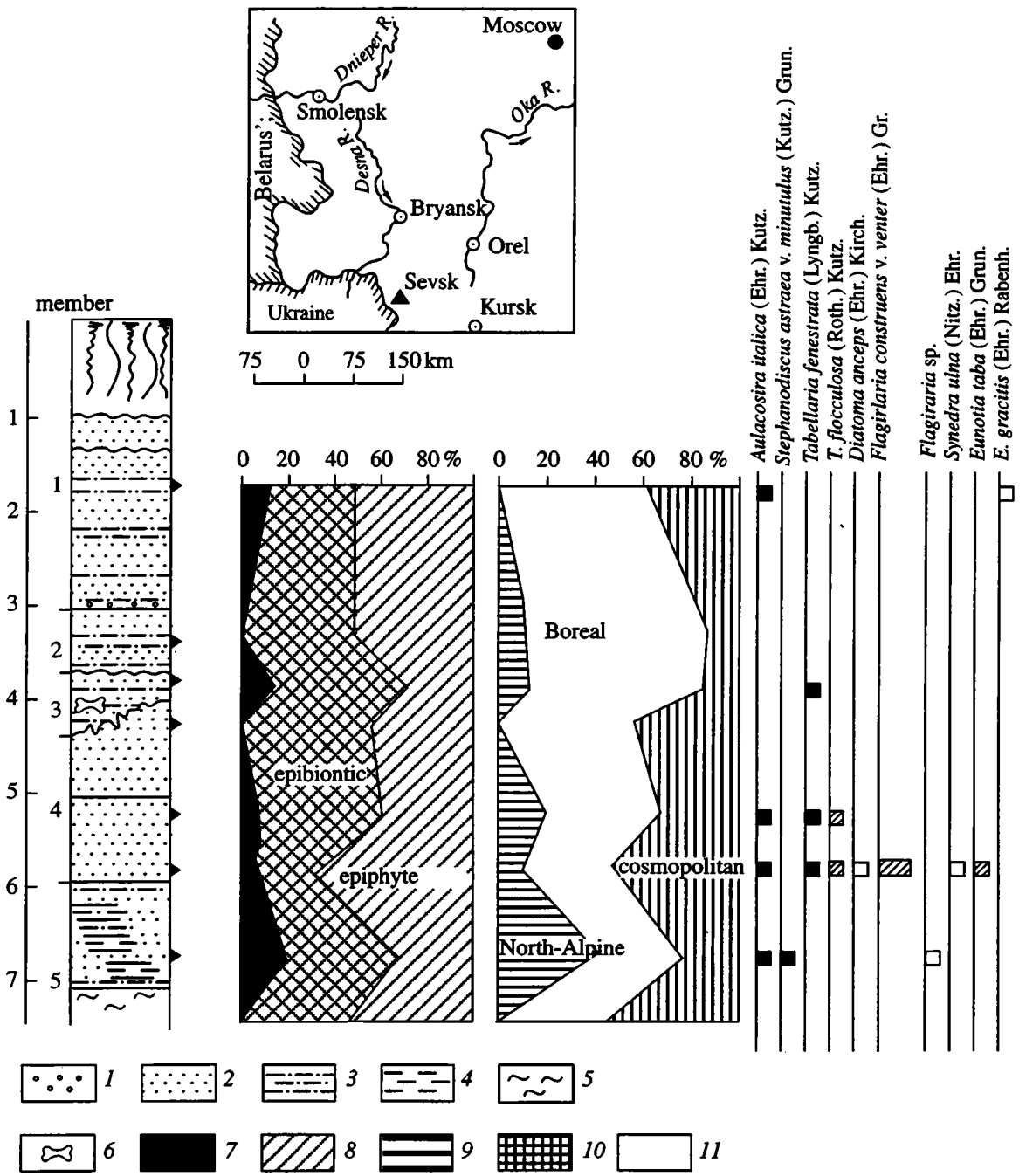
Thus, using the above analysis one may conclude the following.

The diatom assemblage from the depth interval of 5.95 - 7.50 m corresponds to habitat environments in a shallow and quiet water basin of the oxbow lake type. The sediment lithology (dove-colored clays) also favors their lacustrine genesis. Changes in lithology and diatom taxonomic composition up section (member 4, depth interval 5.05 - 5.95 m) show that the depositonal setting was transformed from an oxbow lake into a flood plain. Peculiarities of the diatom assemblage, such as an abundance of benthic forms, especially epiphytes, coexisting with some rheophilic and edaphic elements, strongly indicates flood plane environments (Jouse, 1939). Rather abundant palustrine forms and species of *Caloneis silicula* v. *truncatula*, *Cymbella leptoceras*, and *Gomphonema acuminatum* typical of stagnant basins testify to flood plane bogging. Deposits of this time interval comprising fine-grained, horizontally bedded sands correspond well to the sedimentary facies of the suggested setting.

Further overgrowth of the flood plain seems to be a result of its exsiccation, which caused a sedimentation break, as is evident from the uneven boundary marked by erosion pockets between the bone-bearing and underlying deposits. Changes in diatom assemblages show that, above this surface, a shallow bog lake of formed dystrophic type with an acid water reaction and very low content of plankton, i.e., with features characteristic of a swamp. This lake served as a collector of mammoth carcasses transported by the river stream.

Diatom assemblages from deposits overlying the bone-bearing beds (members 1 and 2) retain features of dystrophic lake flora.

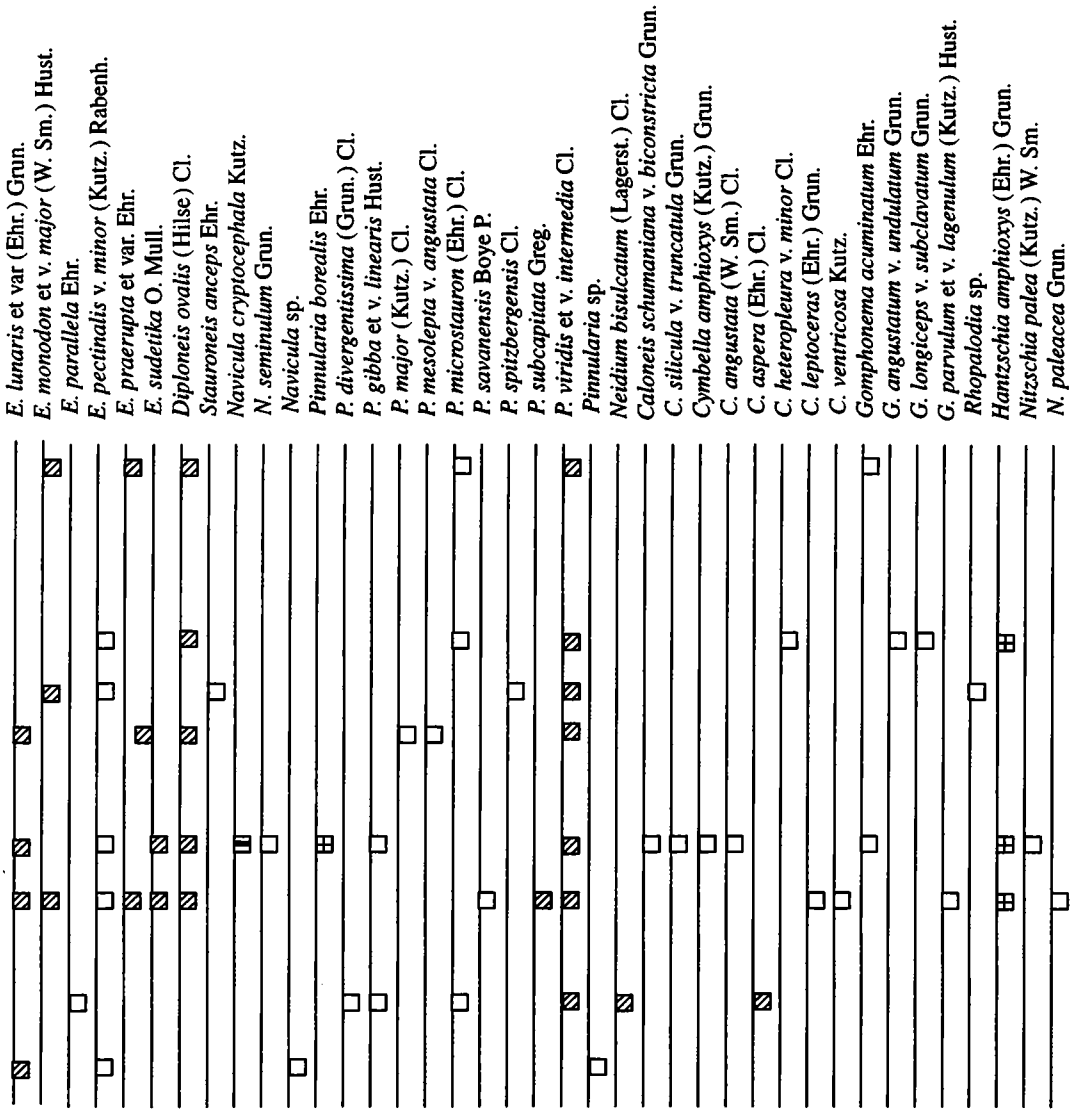
In general, the flora was Boreal, i.e., it was characteristic of temperate latitudes (Khursevich, 1986). Still, an elevated proportion of North-Alpine elements, which is especially high in assemblages from the bone-bearing and underlying deposits, suggest a colder climate in the periglacial zone of the last glaciation than at present.



Sevsk locality map and diagram of diatom distribution in the section of the mammoth burial.
1 – pebbles or gravel; *2* – sand; *3* – sandy loam; *4* – silt; *5* – mud; *6* – bone-bearing lens; *7* – planktonic species; *8* – palustrine and paludo-lacustrine species; *9* – rheophilic species; *10* – edaphic species; *11* – other benthic forms.

A similar trend in the evolution of paleogeographic environments was inferred from data of microtheriological and palynological studies.

As was mentioned, the radiocarbon age determinations are 13 950 ± 70 years (GIN no. 5778) for the bone-bearing beds and 13 680 ± 60 years (GIN no. 6209) for the overlying deposits (Sulerzhitskii and Lavrov, 1993). The datings correspond to the termination of the late Valdai epoch, when cold tundra and steppe provinces characterized the landscape. During the transitional period to the Holocene, a gradual climate warming and related changes in landscape and vegetation resulted in mammoth fauna extinction. These climatic and paleogeographic transformations have been recorded by fossil


E. lunaris et var (Ehr.) Grun.
E. monodon et v. major (W. Sm.) Hust.
E. parallela Ehr.
E. pectinalis v. minor (Kutz.) Rabenh.
E. praerupta et var. Ehr.
E. sudetika O. Mull.
Diploneis ovalis (Hilse) Cl.
Stauroneis anceps Ehr.
Navicula cryptocephala Kutz.
N. seminulum Grun.
Navicula sp.
Pinnularia borealis Ehr.
P. divergentissima (Grun.) Cl.
P. gibba et v. linearis Hust.
P. major (Kutz.) Cl.
P. mesolepta v. angustata Cl.
P. microstauron (Ehr.) Cl.
P. savanensis Boye P.
P. spitzbergensis Cl.
P. subcapitata Greg.
P. viridis et v. intermedia Cl.
Pinnularia sp.
Neidium bisulcatum (Lagerst.) Cl.
Caloneis schumaniana v. biconstricta Grun.
C. silicula v. truncatula Grun.
Cymbella amphioxys (Kutz.) Grun.
C. angustata (W. Sm.) Cl.
C. aspera (Ehr.) Cl.
C. heteropleura v. minor Cl.
C. leptoceras (Ehr.) Grun.
C. ventricosa Kutz.
Gomphonema acuminatum Ehr.
G. angustatum v. undulatum Grun.
G. longiceps v. subclavatum Grun.
G. parvulum et v. lagenulum (Kutz.) Hust.
Rhopalodia sp.
Hantzschia amphioxys (Ehr.) Grun.
Nitzschia palea (Kutz.) W. Sm.
N. paleacea Grun.

(Contd.)

diatoms. The mammoth burial is in the location where a shallow bog lake existed earlier.

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Interdepartmental Conference on General Stratigraphy

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Abstract – The conference that was held January 28 - 30, 1994, in St.Petersburg was organized by the Interdepartmental Stratigraphic Committee of Russia and the All-Russian Institute of Geology. More than 100 specialists participated who work in different institutions of the Russian Committee of Geology and Mining, the Russian Academy of Sciences, and universities and other departments from different cities of the CIS: St. Petersburg, Moscow, Apatity, Ekaterinburg, Krasnodar, Leninsk-Kuznetskii, Nizhnii Novgorod, Novosibirsk, Okha, Perm', Syktyvkar, Tyumen', and Ufa as well as Vil'nyus, Kiev, and Minsk.

The declared subjects of the conference involved a wide range of problems in modern stratigraphy, including the object and subject of the science, principles and criteria of defining the boundaries of local units, stratigraphic classification and the spatial and temporal aspects of stratigraphy, new trends and methods of investigation, as well as stratigraphic coverage of regional geological prospecting.

The first (All-Union) conference with a similar range of topics was held at the All-Russian Institute of Geology in 1955 (*Reshenie* ..., 1955). Two principal papers were presented and discussed: "Stratigraphic and Geochronological Units (Their Principles, Content, Terminology, and Application Rules)" by Librovich, and a paper by Rotai on the stratigraphic classification project. Other papers on various problems were also presented by Menner, Leonov, Zhizhchenko, Bodylevskii, Rauzer-Chernousova, Nalivkin, Keller, Vyalov, Tikhvinskaya, Khalfin, and others. This conference passed a series of resolutions on stratigraphic classification, terminology and nomenclature, stratotypes, stratigraphic and geochronological scales, and the classification suggested by Hedberg. The meeting recommended that laboratories intensify their work on the determination of the absolute ages of rocks; it also recommended the creation of a stratigraphic catalog, the compilation of *Stratigraphic Glossary*, and the founding of *Paleontological Journal*.

The application of the conference to the USSR Ministry of Geology, the Presidium of the USSR Academy of Sciences, the Ministry of Oil Industry, and the Ministry of Higher Education with a request to found the Interdepartmental Stratigraphic Committee (ISC) of the USSR under academician Nalivkin as chairperson was vital. In the same year (1955), ISC was founded to attend to the realization of the resolutions adopted at the conference and to elaborate the stratigraphic basis for the State Geological Map, the compilation of which had started in 1954 at a scale of 1 : 200 000 (Zhamoida and Prozorovskaya, 1985).

By deciding at the beginning of 1992 to convene another stratigraphic conference, ISC intended to sum up the achievements of Russian stratigraphers over the past 40 years, debate the key problems, and elaborate on recommendations for theoretical and practical stratigraphy. However, the events occurring in our country caused the range of problems and number of participants to decrease and made the elaboration of certain recommendations impossible. Nevertheless, the conference appears to have been a success, which can be seen from what follows below.

The conference began with the 7th Karpinsky Memorial Symposium. In his opening speech, Putintsev, Deputy-Director of the All-Russian Institute of Geology, emphasized the importance of the first All-Russian Conference on General Stratigraphy and briefly elucidated further ISC activity. Zhamoida, Chairperson of ISC (1994), delivered the paper "*Startigraphic Space, or the World of Stratigraphy*."

Twenty-two presentations were submitted, which were divided into four groups: (1) Stratigraphy and Its Structure: Boundaries of Local Units; (2) Precambrian Stratigraphy; (3) New Methods of Stratigraphy; and (4) Stratigraphic Practice. The first group incorporated presentations made by Zhamoida, Egoyan, Tikhomirov, Chernykh and Chuvashov, Krasnov, Rostovtsev, Kagarmanov, and Nenashev, Tsyganko, and Solov'eva.

In his report "*Stratigraphy as a Description of the Earth's Layers*," Egoyan (Termneft' Krasnodar Institute) considers this science to be based exclusively on the description of the facts, which attaches a particular significance to the description of local units representing volumetric geological bodies within a certain region. All stratigraphic inferences should be based on the facts collected from sections. Various stratigraphic categories, scales, "gold nails," etc., imply only the

interpretation of facts. He believes that the Russian Stratigraphic Code and, thus, other national codes now in operation should be revised. During the discussion, Egoyan presented his own determinations of principal stratigraphic notions and stressed the necessity of analyzing stratigraphic errors.

Tikhomirov (Moscow Institute of Geological Prospecting) considers stratigraphy to be the science of the rythmic regularities in the stratisphere structure, which reflect stages of different ranges in the evolution of the planet and the organic world. The local units boundaries are indications of pulsating vertical changes in the world's sea-level with due regard for changes in biota.

Chernykh and Chuvashov (Institute of Geology and Geochemistry, Ural Division, Russian Academy of Sciences), in their presentation "*Stratigraphic and Geochronological Units*," emphasize the necessity of clearly distinguishing between the notions of a *geochronologic unit* (*chronon*) and a *lithostratigraphic unit* (*straton*), the criteria for determining their boundaries being substantially different. Another communication by Chuvashov was dedicated to litho- and biostratigraphic boundaries in the stratotype section of the Permian System, their correlation, and the importance of stratotype. The author subdivides the biostratigraphic boundaries into ecostratigraphic and phylogenetic ones.

The stratigraphic classification structure was considered by Krasnov and Rostovtsev.

Krasnov (Siberian Research Institute of Geology, Geophysics, and Mineral Resources) believes that it is unreasonable to use a multiplicity of features for the elaboration of any classification. Thus, he supports the idea of singling out those criteria of local units that are based on one feature, which, according to the author, is genetic. Main stratigraphic units are represented by biostratigraphic (planetary and correlative) and lithostratigraphic (local) units.

Rostovtsev (All-Russian Institute of Geology) suggests two classes of stratigraphic units: lithologic, which is characterized by the lithology of the units, and lithologic-evolutionary, which incorporates those units singled out on the basis of the study of the organic world's irreversible evolution.

Tsyganko (Institute of Geology, Komi Research Center, Ural Division, Russian Academy of Sciences) distinguishes three types of boundaries of stratigraphic units: stratigraphic boundaries proper (isochronous and potentially planetary), parastratigraphic boundaries of regional units in continuous sequences (impersistent and diachronous), and allostratigraphic boundaries (of regional units in sequences with hiatuses).

In the abstracts to her paper "*On Principles for Subdivision of Local Units and Criteria for Defining Stratigraphic Boundaries*," Solov'eva (Geological Institute, Russian Academy of Sciences) argues that only the initially determined volume of a stratigraphic unit may dictate its boundaries, and that the principal task of stratigraphy at the present stage involves the correlation of regional units with those of the conventional general scale.

Four communications were devoted to different problems of Precambrian stratigraphy.

Bekker (All-Russian Institute of Geology) in his report "*Chronometric Systems: Illusion or Reality of the Precambrian?*" considered the relationship of stratigraphic and geochronologic units of the Precambrian general scale. In contrast to the Phanerozoic where the units are strictly ranked, the units in the Precambrian are more liberal. The *Precambrian Scale* suggested by the International Commission on Stratigraphy comprises 10 periods in the Proterozoic, whose boundaries and volume may change greatly in the course of distinguishing geological systems in the future because of the lack of stratotypes. Criteria for the subdivision of Proterozoic stratigraphic units of this rank were enunciated using the terminal Precambrian System as an example.

Negrutsa and Negrutsa (Institute of Geology, Kola Research Center, Russian Academy of Sciences) emphasized the particular significance of local units of the Precambrian and the role of stratotypes. Imperfections of the general, regional, and interregional scales of the Precambrian result from imperfections in the compilation and correlation of local stratigraphic schemes. The authors believe that the principles and criteria used for Phanerozoic stratigraphy are quite applicable to the Precambrian. In his report "*On the Precambrian Cyclorhythmostratigraphy*," Negrutsa recognizes cycles and rhythmicality as the most shining characteristics of the Precambrian structure. He suggested a compound cyclorhythmometric scale of the Earth's history, which enables the role of hiatuses in stratigraphy to be determined.

Bogdanov and Shul'diner (All-Russian Institute of Geology) placed particular emphasis on the necessity of improving the Precambrian regional stratigraphic scales, since these help in correlating local sections and linking them with the general scale. They proposed that synthems, serials (2 - 3 series), and horizons should be singled out in the Precambrian but not in the complexes.

Certain methods of stratigraphy were elucidated in six of the presented papers.

In his paper "*Transition to Infrazonal Subdivision of Sedimentary Sequences: A Call of the Time*," Gladenkov (Geological Institute, Russian Academy of Sciences) demonstrated, using Neogene deposits as an example, the possibility of singling out fine units (zone, subzone, layer, cyclic unit, and member), whose time-intervals are similar to those of the Quaternary units.

Using the example of the Devonian and Carboniferous of the Urals, Shcherbakov, Shcherbakova, Kochneva, Stepanova, and Kucheva (Perm' State Technical University) argue for the superfine subdivision of deposits and their correlation on the basis of the biorhythmostratigraphic method, which involves the combination of biostratigraphy with a cyclic analysis of

sedimentary sequences when relating cyclic units to biostratigraphic boundaries. The regional importance of cyclic units was shown as well.

Makhlina (Central Production Geological Association) examined in her report biostratigraphic boundaries and the criteria for their determination. She emphasized that the distinguishing and tracing of rhythmic lithostratigraphic units with various facies is the most important task when elaborating and substantiating regional and local stratigraphic schemes.

Akhmet'ev (Geological Institute, Russian Academy of Sciences) reported on the problems of phytostratigraphy, the criteria for distinguishing phytochrons, and the causes for their succession.

Sal'nikov (Sakhalin Research and Design Institute for Oil and Gas Industry) devoted his paper about local units to the problem of the different ages of their boundaries, which makes structural and other reconstructions for conducting prospecting work difficult, but does not inhibit the mapping of these units at any scales. He assigns a layer to main local units.

The problems of sequence stratigraphy, which is currently popular, were considered in the paper presented by Dronov and Savitskii (St. Petersburg University). They showed the effectiveness of the method of subdiving the sedimentation basin cover into sequences.

Papers closing the conference were devoted to stratigraphic practice. The problems considered are as follows: the implications of substantiated detailed stratigraphic schemes, reference legends, and the fulfillment of the requirements for distinguishing formations (Shik, Central Production Geological Association); the necessity of indicating in the map legend the limits of changes in the stratigraphic volume of formations (Kagarmanov, St. Petersburg Mining Institute; Nenashev, All-Russian Institute of Geology); the ambiguity of the notion of a *geological region*, and the preferability of the elaboration of stratigraphic schemes for stratigraphic regions (Kagarmanov). Prozorovskii (St. Petersburg University) emphasized the necessity of compiling more illustrative, concrete, and space-saving schemes. He also proposed that the form of their submission should be changed: the correlative part should depict the peculiarities in the spatial distribution of local units, the unified part should be composed of regional biostratigraphic and consistent lithostratigraphic units, and all the descriptive information should be placed in the explanatory note.

Borisov (All-Russian Institute of Geology) spoke about the difficulties that arise when correlating local and general units as well as how to overcome them, using the Cenozoic continental sequences of the Altai–Sayan area as an example.

Some problems were touched on by participants in the final session.

Gladenkov proposed that similar meetings should be held in the future. He also stressed the implications of certain trends of investigations in modern stratigraphy.

Rublev called attention to the necessity of choosing reliable reference points in order to improve the general isotopic-geochronologic scale, because only 8 such points now exist for the entire Phanerozoic. Shkatova showed the feasibility of applying the paleomagnetic method to control age determinations. L.N. Dorofeeva reported on the application of the method, elaborated in the All-Russian Institute of Geology, of a carbonate thermometer together with the method of an oxygen thermometer to the subdivision and correlation of deposits. Semikhatov considered the essence of chronostratigraphy, the determination of the lower boundaries of general units with the help of global events, and land criteria of synchronous and isochronous characters of different types of stratigraphic boundaries. Preobrazhenskii, Golovenok, Nikol'skii, Bogdanov, Negrutsa, Negrutsa, Lezhnin, Krasnov, and other scientists took part in the discussion.

In his concluding speech, Zhamoida appraised the results of the conference, then enlarged upon the relationships of a unit volume and its boundaries and a geological body and its section. He also considered the necessity of detailed subdivision of deposits and, at the same time, the negative aspects of redundant separation of zones or cyclic units as well as the necessity of analyzing the startigraphic errors. He discussed some of the remarks and proposals for the *Stratigraphic Code* that had been quoted in some papers.

After the discussion, the conference emphasized the importance of the second issue of the *Stratigraphic Code of Russia* for solving the pressing problems of general stratigraphy, achieving uniformity in the interpretation of principal stratigraphic notions and procedures, and improving the classification of stratigraphic units. It was also stressed that, recently, Russian stratigraphers has been paying most of their attention to stratigraphic unit boundaries; their complex (event, ecosystem, magnitostratigraphic, seismostratigraphic, and, in some cases, hemostratigraphic) substantiation; and the application of a wide range of paleontological, historical-geological, and physical methods for their correlation.

However, some aspects of general stratigraphy require further, detailed elaboration and wide discussion. In the participants' opinion, these are: principal notions used in stratigraphy, including the object and subject of research; the problem of so-called lithodemic units, which is especially important for the Precambrian; the classification of stratigraphic units of principal categories with due regard for new special categories; the relationship of stratigraphic units of different categories and geologic time, and improvement of the general geochronological scale; the classification of boundaries of stratigraphic units; refinement of the general (international) stratigraphic scale; a complex

substantiation for the determination and correlation of stratigraphic boundaries of general units, with the selection of global stratotypes of the boundaries (*Stratigraficheskii Kodeks*, 1992); improvement, including detailing, of regional and local stratigraphic schemes, with due regard for paleogeographic and (or) paleobiogeographic zoning (basin stratigraphy).

The meeting came to the following conclusions:

(1) The Bureau of the Interdepartmental Stratigraphic Committee should apply to the Committee on Geology and Mineral Resources of the Russian Federation and the Russian Academy of Sciences with a request to encourage stratigraphic investigations that serve as a basis for geological surveying, prospecting, and exploration as well as general geologic knowledge, and that steps should be taken to ensure the generalization and storage (using computers) of gathered stratigraphic materials and the elaboration of new effective methods for the subdivision and correlation of local units and their boundaries. Also noteworthy are the acute necessity to provide valuable and advanced equipment for stratigraphic works and the indispensability of organizing interdepartmental regional stratigraphic meetings in future.

(2) Interdepartmental conferences on general stratigraphy should be held regularly, and the principal theme, or even 2 to 3 themes, should be determined in advance for the discussion and adoption of the corresponding resolutions. The Interdepartmental Stratigraphic Committee should become the organizer of these meetings and hold them at least once every three years.

(3) To apply to the Editorial Board of the journal *Stratigraphy and Geological Correlation* with a request to arrange a section in the journal entitled *Symposium by Correspondence* for discussion of the cardinal problems of general stratigraphy.

4. Taking into account the significance of correct isotopic age determination of stratigraphic unit boundaries for the general scale of the Phanerozoic, the conference highlights the necessity of revising the currently available geological scales and believes that, for this purpose, it is urgent that the work of the Isotopic-Geochronological Commission of ISC be resumed.

5. Participants of the conference are requested to send their presented papers to the Editorial Board of *Stratigraphy and Geological Correlation* for publication, including within them concrete proposals for stratigraphic classification and terminology and stratigraphic notions. They are also requested to send suggestions for the organization of stratigraphic service in modern conditions to the Bureau of ISC.

All the participants unanimously acknowledged that the conference was useful and essential.

Reviewer B.S. Sokolov

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